Lake status records from China: Data Base Documentation

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1. Introduction

This report documents reconstructions of long-term changes in lake status from 42 lakes from China. The present version of the Chinese Lake Status Data Base (CLSDB.1, February 2001) is part of an ongoing international effort to produce a new global lake data base (Kohfeld and Harrison, 2000) designed to be used to validate climate models. The structure of the CLSDB therefore parallels other regional data bases that have been or are being compiled under the auspices of this international effort, such as the European Lake Status Data Base (Yu and Harrison, 1995), the FSU and Mongolia Lake Status Data Base (Tarasov et al., 1994; 1996), the North American Lake Status Data Base (Harrison et al., subm.) and the North African Lake Status Data Base (Hoelzmann et al., in prep).

1.1. Lakes as Indicators of Past Climate Changes

Lakes respond to changes in the local water budget (precipitation minus evaporation over the lake surface and its catchment) by changing in depth and area. On a Late Quaternary time scale, lake status (a qualitative index of changes in lake level, area or relative water depth) is in equilibrium with the changing climate (Street-Perrott and Harrison, 1985; Harrison and Digerfeldt, 1993; Cheddadi et al., 1996; Harrison et al., in press) and therefore can be used as an indicator of past climate. Both closed-basin and overflowing lakes have been used successfully to reconstruct past climate changes (e.g. Street-Perrott and Harrison, 1985; Digerfeldt, 1986; Cheddadi et al., 1996; Harrison et al., in press). Syntheses of lake status data have become an important tool for reconstructing palaeoclimatic changes at a continental to global scale (e.g. Street-Perrott and Harrison, 1985; COHMAP Members, 1988; Street-Perrott et al., 1989; Harrison et al., 1996; Kohfeld and Harrison, 2000).

1.2. Chinese Lakes as Indicators of Asian Monsoonal Climate Changes

The modern climate of China (15-56°N, 60-140°E) is largely determined by the interplay of summer and winter monsoonal circulation (Fig. 1). The winter monsoon associated with the Siberian high pressure system governs almost the whole of western and central China, and brings cold and dry continental air southward to ca 20°N (Ren et al., 1979). The influence of the winter monsoon is also felt further south than this, where it becomes associated with the so-called Plateau Monsoon (Sheng CY, 1986) of the Tibetan Plateau. The summer monsoon associated with the Pacific subtropical anticyclone brings warm and moist marine airmasses from the western Pacific Ocean to eastern China as far north as 50°N, while the Indian monsoon circulation brings moist marine air from the Indian Ocean to southern Tibet and southern China (Chinese Academy of Science, 1985). Changes in the intensity and range of the Asian monsoons produce changes in regional temperature and precipitation patterns across China (Fig. 2). Thus, the penetration of the Pacific monsoon produces summer (June, July, August: JJA) precipitation ranging from >750 mm (in the south) to >250 mm (in the north) along the coastal regions of eastern China. A similar gradient is seen in winter (December, January, February: DJF) rainfall, with >200mm in the southeast and <50 mm in the northeast. The winter rainfall in the eastern coastal region is associated with frontal disturbances. The penetration of the Indian monsoon results in high summer rainfall on the central and eastern parts of the Tibetan Plateau and on the Yunnan Plateau (ranging from >250 mm to >750 mm). The eastern Tibetan Plateau, which is partially sheltered from the influence of the Indian monsoon and which is influenced by cold, dry winds associated with the Asian winter monsoon, is extremely dry. Summer precipitation values are <100mm, while the winter rainfall is <10mm. Northwestern China (inland Xinjiang) is dominated by continental westerly winds (associated with the upper level Westerly jet) in summer and lies beyond the limits of both the Pacific and the Indian monsoons. As a result, summer precipitation is low (<100mm). These regions are dominated by the Asian winter monsoon, and winter rainfall is therefore also generally low although westerly systems bring some (< 50mm) rain to the northernmost part of Xinjiang. Millenial-scale changes in the strength of the summer and winter monsoons are registered by a variety of palaeoenvironmental records (e.g. Shi et al., 1993), including the geomorphic and biostratigraphic records of lakes.

There are 487 natural lakes with an area of more than 10 km^2 in China (Wang and Dou, 1998). Freshwater lakes (183 lakes), are mostly distributed in the eastern part of China (154 lakes). Saline and hypersaline water lakes (ca 276) mainly occur in the western part of China (Fig. 3) (Note: Salinity information is not available for the remaining 28 lakes). The distribution of fresh and saline lakes is consistent with observed long-term patterns of P-E (Chinese Academy of Science, 1981; Lu and Gao, 1984). Most freshwater lakes are found where there is a positive water balance (P>E) (Fig. 3) and mean annual precipitation is >600mm. Where the water balance is negative (P<E) and mean annual precipitation is <600mm, saline lakes predominate. Thus, the freshwater lakes are mostly located in regions under the influence of the Pacific and/or Indian summer monsoons, while saline lakes are distributed in regions uninfluenced by the Pacific and/or Indian summer monsoons.

Figure 1: Regions of China and the limits of monsoonal circulation

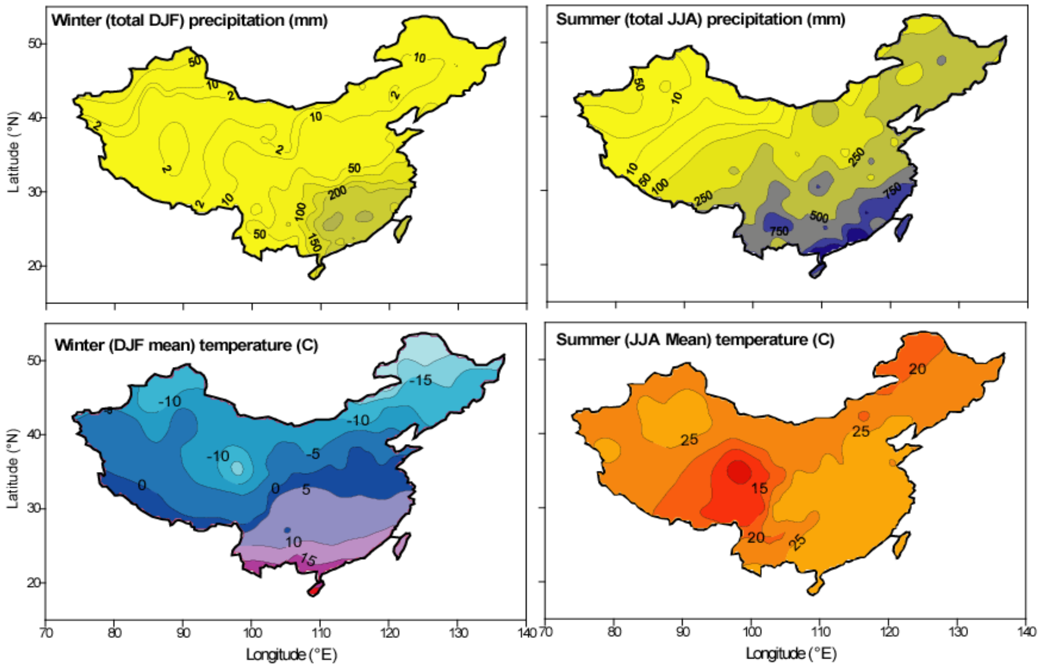

Figure 2: Winter (December, January, February: DJF) and summer (June, July, August: JJA) temperature and precipitation patterns over China

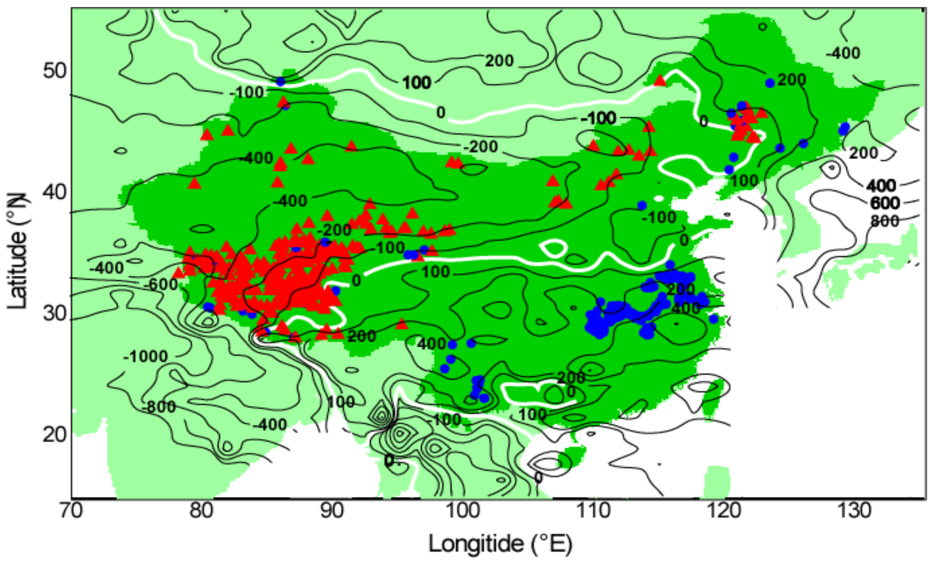

Figure 3: The distribution of saline (red triangles) and freshwater (blue circles) lakes across China in relation to mean annual P-E (mm). The 459 lakes plotted in the figure all have areas greater than 10 km^2. The white line marks the boundary where P-E = 0 mm.

1.3. Previous Work on Palaeohydrological Changes in China

Investigations of Chinese lakes began several decades ago (Chinese Geology Society, 1956). Although some of these studies documented changes in lake level in response to climate changes (e.g. Hedin, 1922; Pei and Li, 1964), most studies focussed on mineral or water resources (e.g. Zheng MP et al. 1989, Zheng XY et al., 1992, Zhu et al. 1989) or reconstructions of regional vegetation changes (Huang CX et al., 1983; Du and Kong, 1983).

The recognition that lake data could be used to reconstruct past regional climate changes led to the construction of data bases. The Oxford Lake-Level Data Base (OLLDB) was completed in the late 1980's (Street-Perrott et al., 1989). The OLLDB is global in scope but only contains information about closed lake basins and the lake-level changes are coded for 1000 year intervals. There are only nine sites from China in the OLLDB (Table 1). Jin-qi Fang (1991) collected data from 122 lakes from China and used these data to plot lake-level maps for the Last Glacial Maximum and the Holocene. Unfortunately, 30 % (42 sites) of the sites in this data synthesis have no palaeorecord (Table 2). Furthermore, an additional 22 % of the data have a palaeorecord for only a single time period, such that it is impossible to derive a record of changes in lake status relative to today. The absence of documentation for the coding or of the chronologies of the remaining records makes it difficult to determine whether these records are compatible with lake codings from the OLLDB or the regional lake data bases for e.g. Europe (Yu and Harrison, 1995) and Russia (Tarasov et al., 1994, 1996).

Table 1. Chinese lakes in the OLLDB (Street-Perrott et al., 1989)

BMN	Site	Lat. (N)	Long. (E)	Elev. (m a.s.l.)	Record length (ka B.P.)
6203	Chenco	29.00	90.70	4000	0-6
6204	Lop Nor	40.50	90.50	780	0-2
6319	Da Qaidam	38.00	95.20	2900	0-21
6320	Kunteyi	38.50	92.30	2790	0-15
6321	Qaidam	37.50	94.50	2700	0-30
6322	Qinghai	36.80	100.50	3266	0-11
6323	Xiao Qaidam	37.40	95.40	2900	0-15
6324	Zhangzang Chaka	32.50	82.33	6000	0-16
6325	Zharinanmuco	31.08	85.55	4613	0-7

There have been a large number of investigations on Chinese lakes in the last two decades. These investigations have made use of biological evidence (e.g. diatoms: Li WY et al., 1983; ostracodes: Huang BR, 1990; molluscs: Dong et al., 1983), shoreline information (Qui SW et al., 1988; Mehmet I, 1988; Li BY et al., 1991; Li SK, 1992), geochemistry (Geng and Cheng, 1990; Zheng XY et al., 1992; Li BX et al., 1989; Zheng MP et al., 1989), stable isotope (Han and Yuan 1990; Gu ZY et al., 1993; Wu JL et al., 1993) or multi-proxy approaches to reconstruct palaeohydrological and palaeoenvironmental changes, and have emphasised the use of radiometric methods for chronologies (Qiu SW, 1988; Zheng HH, 1989; Li XG et al., 1984; Gu et al., 1990; Qi and Zheng, 1995; Huang Q et al., 1980; Li BY et al., 1983). The increase in the number of reasonably well-dated sites for which data is available makes it timely to attempt a new synthesis of Chinese lakes data, using documentation, interpretation and coding techniques compatible with the GLSDB. Since much of the work on lakes has only been published in Chinese, the new synthesis serves the additional function of making these palaeohydrological studies available to the wider scientific community.

Table 2. Chinese lakes in Fang (1991)

Lake	Lat. (N)	Long. (E)	No. of coded periods	Note
AKSAYQIN	35.20	79.83	30	
ANGLARU	29.10	87.15	0	no coding
BAILINGHAI	37.90	97.00	2	
BAIMAHU	31.60	119.10	1	coding only for 0ka
BAIXIAN	31.12	120.80	1	
BAIYANGDIAN	38.31	116.10	7	
BAIYUN	36.90	117.40	1	coding only for 0ka
BAMOCUO	31.25	90.58	0	no coding
BANGONGCUO	33.73	79.83	12	
BANGUOCUO	31.85	89.30	1	
BANGDACUO	34.90	81.53	3	
BAOYING	33.20	119.20	1	coding only for 0ka
BEIHU	30.60	120.20	1	coding only for 0ka
BELL	47.90	117.8	1	coding only for 0ka
BIEHUA	36.90	119.00	0	no coding
BOSHITENG	42.37	87.07	0	no coding
CAOHAI	26.60	100.20	18	
CHANGMAO	29.90	120.65	0	no coding
CHARHAN	35.52	93.85	31	
CHEN	31.20	120.83	1	
CIBI	26.10	99.93	1	
DABUXUN	37.00	95.05	20	
DACHAIDAN	37.50	95.23	26	
DAIHAI	40.37	112.40	11	
DAISHAN	35.00	116.00	1	
DAJIUHU	31.80	110.70	5	
DALU	37.30	114.80	1	
DAMAO	29.85	120.70	0	no coding
DANHU	31.30	120.55	1	
DANGHU	30.80	121.50	0	no coding
DANYANG	31.40	118.70	3	
DESHENG	32.95	119.92	0	no coding
DIANCHI	24.85	102.70	31	
DONGPING	35.98	116.20	1	coding only for 0ka
DONGTING	29.20	112.50	7	
DONGZE	36.00	114.40	0	no coding
DOUSIKULEI	38.01	91.00	4	
DRYLAKE3	36.80	95.20	21	
DUOGINGCUO	27.00	89.20	1	
DUSHAN	35.05	116.80	0	no coding
ERHAI	25.79	100.20	28	
FULEN	48.57	117.23	11	
FULUN	48.95	117.40	1	coding only for 0ka
FUXIAN	24.50	102.87	9	
FUYONG	31.50	120.50	1	
FUZE	35.45	112.25	2	
GAOJIZE	37.03	116.00	0	no coding
GAOYU	32.90	119.20	2	
GUCHEN	31.25	118.92	1	
GUOPEIYAN	31.60	119.9	1	
HEZE	35.10	115.85	0	no coding
HONGZE	33.40	118.55	1	coding only for 0ka
HONGHU	29.70	113.20	3	
HUANGZHANG	40.00	118.50	2	
JIANFUZE	34.80	114.15	1	
JILEIER	37.50	94.00	5	
JINXINGZE	35.00	110.38	0	no coding
JUDIAN	37.00	118.65	2	
JUYANZE	42.40	100.70	4	
KUNCHEN	31.57	120.74	3	
LEIXIA	35.45	115.60	1	
LIANGCHEN	31.45	119.60	1	
LIANHU	32.10	119.55	1	
LIULIPO	24.85	118.60	1	coding only for 0ka
LONGGONG	22.17	100.40	9	
LOP	40.20	90.15	17	
LUO	40.50	91.60	2	
LUOMA	34.10	118.20	22	
MANDONGCUO	33.51	78.95	7	
MANYANG	26.20	100.40	4	
MAODUDIAN	35.70	116.30	1	coding only for 0ka
MENGZHUZE	34.65	115.75	1	
NANSHI	35.00	116.95	1	coding only for 0ka
NANWANG	35.70	116.10	1	coding only for 0ka
NANYANG	35.30	116.60	0	no coding
NARIYONG	28.30	91.95	4	
NORTHTIANSHUI	35.50	79.58	5	
PEIKUCUO	28.80	85.45	2	
PINGZHOUCHEN	37.04	118.50	0	no coding
POYANG	29.05	116.20	15	
PUTIAN	34.80	114.50	3	
QILIHAI	39.30	117.50	5	
QINGHAI	36.80	100.18	31	
QINGSHA	37.00	118.00	1	coding only for 0ka
SHANGNAN	30.50	120.30	0	no coding
SHAOYANG	35.00	116.90	1	coding only for 0ka
SHAOYUQISOU	37.20	112.00	1	
SHEIYANG	33.30	119.60	3	
SHENGPO	35.60	111.45	1	
SHIGUANG	32.50	119.50	1	coding only for 0ka
SHIJUIHU	31.45	118.90	2	
SHIPANGUR	33.33	80.66	1	
SHEYANZE	37.85	108.40	19	
SHUOPO	32.30	116.70	1	coding only for 0ka
SHUOXIANG	34.15	119.10	1	coding only for 0ka
TAIBAI	30.00	115.80	1	coding only for 0ka
TAIHU	31.12	120.10	3	
TIANSHUIHAI	35.35	79.50	6	
TIANYANG	20.50	110.30	5	
TAITEIMA	36.85	88.10	1	coding only for 0ka
TUSHEN	39.20	106.60	2	
WEISHAN	34.50	117.10	0	no coding
WEN'ANWA	39.00	116.20	1	
XIANANHU	30.45	120.30	0	no coding
XIANGHU	30.10	120.18	2	
XIHU	31.25	120.12	2	
XUSHANPO	36.80	117.60	1	coding only for 0ka
YANGCHEN	31.40	120.75	1	
YANGHU	31.50	119.70	1	
YANGYUZE	34.60	109.90	1	
YI'ANTAN	36.60	119.90	1	
YILIPING	37.45	93.40	9	
YINGZE	34.85	113.50	1	
YISIKUR	37.20	78.60	1	coding only for 0ka
YONGHU	31.60	119.80	1	
YUNMENG	29.80	113.20	2	
YUPU	30.00	120.15	2	
YUSHAO	31.45	120.55	1	
Z.CAKA	32.60	82.37	14	
ZHANGZE	35.00	110.30	0	no coding
ZUOERHU	39.85	76.60	1	coding only for 0ka
ZHEHU	30.90	121.70	0	no coding

1.4. Data and Methods

1.4.1. The Data Set

The present data base contains fully documented and coded records from 42 sites across China. The data base was compiled by extracting information on changes in lake status from existing geomorphological, lithological, geochemical, or biostratigraphical data. Most of the primary data were derived from Chinese publications between 1978-1998 (see Reference Section), but additional data was provided by scientists working on individual sites.

Most of the sites have records extending back to the late glacial and/or the last glacial maximum records, and over half of the lakes have records beyond 30,000 yr B.P. A complete documentation exists for every site, describing the primary data, basin characteristics and the interpretation logic.

The standard practice in the GLSDB is that individual modern lakes are only coded as separate sites when they have been hydrologically independent throughout their history. Thus, two modern lakes that lie within a single hydrological basin and are, in fact, relicts of a former expanded lake which occupied that basin at, say, the last glacial maximum, would for the purposes of status coding be treated as a single site. In regions where lacustrine sedimentation may have been quasi-continuous over several glacial-interglacial cycles, and where the possibility of significant changes in catchment area or fluvial networks is therefore large, it can be meaningless or difficult to define sites in terms of hydrological independence throughout the whole of the lake's history. In practice, the amalgamation of individual lakes into a single site has tended to be applied only when the lakes have been hydrologically linked within the last ca 30,000 years (i.e. within the period that is the main temporal focus of the GLSDB). In China, the use of the hydrological independence criterion poses a particular problem. The interval between ca 35,000 and 45,000 yr B.P. was characterised by the existence of several mega-palaeolakes, with areas e.g. of 970 km^2 in Palaeo-Banggongcuo (Li BY, 2000), 1381 km^2 in Palaeo-Zabuye Lake (Zheng MP et al., 1989, 1996), 2650 km^2 in Palaeo-Akesaiqin-Tianshuihai Lake (Wang FB et al., 1990), 16,200 km^2 in Palaeo-Baijian Lake (Pachur et al., 1995), and 25,000 km^2 in Palaeo-Chaerhan Salt Lake (Zheng MP et al., 1996). These lakes were reduced considerably in size after ca 35,000 yr B.P., giving rise to hydrologically independent basins which have remained separate since then. The Palaeo-Akesaiqin-Tianshuihai Lake for example comprised the modern basins of Guozacuo, Akesaiqin, Tianshuihai, North Tianshuihai and Kushiuhai Lakes, and the Palaeo-Banggongcuo comprised the modern basins of Changmucuo, Abucuo and Aiyongcuo. Since the main temporal focus of this compilation is the interval between 30,000 yr B.P. and the present, it would be meaningless to treat the mega-lake basins as other than clusters of individual sites. However, the relationship between the modern (coded) lake sites and the earlier mega-lakes that once occupied these lake basins is fully documented (see Section 3).

The sites chosen for inclusion in the data base have been screened to meet standards of dating control and consistency among different hydrological indicators. Records or parts of records where water depth appears to have been influenced by non-climatic factors, such as tectonism, earthquakes, hydroseral development, changes in the fluvial network or human impact, or by factors where the climatic influence is indirect, such as sea-level changes or glacier fluctuations, have been excluded from the data base.

1.4.2. Sources of Evidence for Changes in Lake Status

The major source of information on changes in lake status for most of the basins discussed in this report are changes in the nature of the sediments and sedimentary structure, as revealed in sediment cores. Palaeoecological evidence (specifically changes in the species assemblages of aquatic plants, diatoms, ostracodes and/or molluscs) has been used to provide corroborating evidence for the reconstructed changes in water depth. Geomorphological evidence, archaeological evidence and historical records have also been used; these lines of evidence often permit reconstruction of absolute lake level. The reconstruction of changes in water depth at every site is based on the consensus interpretation of a minimum of two lines of evidence, following Harrison and Digerfeldt (1993), and the standard practice in the GLSDB.

Changes in water salinity can also be used as an indicator of changes in relative water depth and hence lake status. Qualitative estimates of salinity (fresh, brackish, saline) can be made based on sediment mineralogy and geochemistry, as well as from the diatom, ostracode or algal assemblages. Thus, as the evaporative concentration of the lake water increases, so there is a shift in the mineralogy/geochemistry of the sediments. The general sequence is from carbonate, borate, mirabilite, nitrate to halite with increasing water salinity (Zheng et al., 1989; Zheng et al., 1992). Studies of the aquatic communities in modern Chinese lakes provide the basis to infer water salinity from species assemblages (e.g. Li BY et al., 1994; Li SF, 1996). Tables 3, 4 and 5 summarise the information available in the Chinese literature on the water salinity and depth ranges of diatoms, ostracodes and algae.

Palaeolake shorelines and other geomorphological evidence, documented in the original investigations of individual basins, can be used to indicate absolute lake level and palaeolake area at specific times. When the shorelines are continuous or widespread, the reconstruction of palaeokae area is relatively straightforward. When only isolated shoreline fragments remain, we have assumed that they mark lake level and have used topographic maps to calculate the area of the lake corresponding to this elevation. The presence of shorelines above/below modern lake levels, provided it is consistent with the other lines of evidence about changing water depth available, has been used to reconstruct changes in lake status. In addition, when the lake area for a specific period can be estimated from shoreline or other geomorphic data, these estimates have been incorporated in the database. Such quantitative estimates should provide a strong constraint for evaluation of palaeoclimate simulations using hydrological models (e.g. HYDRA: Coe, 1998) as a diagnostic tool.

Table 3. Water depth and salinity indicated by diatoms (after Hu, 1987; Qi, 1995)

Diatom	Diatom ecology			Water salinity	
	Planktonic	Epiphytic	Benthic	Fresh	Brackish
Achnanthes exigua		yes	yes	yes	
A. lanceolata		yes		yes	
Amphora ajajensis		yes		yes	
A. delicatissima		yes	yes	yes	yes
A. mexicana		yes			
A. ovalis				yes	yes
A. perpusilla		yes		yes	
Anomoeoneis sphaerophora		yes	yes	yes	yes
Aulacoseira granulata	yes	yes			
Cocconeis placentula				yes	
C. disculus		yes		yes	
Coscinodiscus lacustris	yes	yes		yes	
Cyclostephanos dubis	yes				
Cyclotella comta	yes			yes	
C. ocellata	yes		yes	yes	yes
C. commensis	yes			yes	
C. meneghiana	yes		yes	yes	yes
C. kutzingiana			yes		yes
C. quadrijuncta	yes			yes	yes
C. stelligera	yes		yes	yes	yes
C. spp.	yes			yes	yes
Cymbella affinis			yes	yes	
C. minuta		yes		yes	
C. cymbiformis		yes	yes	yes	
C.cistula		yes	yes	yes	
C. gracilis		yes	yes	yes	
C. hustedtii		yes		yes	yes
C. parva		yes	yes	yes	yes
C. helvetia		yes	yes	yes	
C. lanceolata		yes	yes	yes	yes
C. leptocera		yes		yes	
C. lata		yes		yes	yes
C. stelligera	yes	yes		yes	
C. turgida				yes	
C. ventricosa		yes		yes	yes
C. aequalis		yes		yes	
C. lacustris		yes		yes	
C. amphicephala		yes		yes	
C. laevis		yes		yes	yes
Cymato pleura	yes	yes		yes	yes
C. solea	yes		yes	yes	yes
Diploneis elliptica	yes		yes	yes	yes
D. parma			yes	yes	yes
Epithemia intermedia				yes	
E. sorex		yes	yes	yes	
E. zebra		yes	yes	yes	
Eunotia arcus		yes	yes	yes	
E. argus				yes	
E. zebra		yes		yes	yes
E. sorex		yes	yes	yes	yes
E. turgida		yes		yes	yes
E. adnata		yes	yes	yes	
E. reicheltii		yes		yes	
E. proboscidea		yes		yes	
Fragilaria lapponica		yes	yes	yes	yes
F. brevistriata			yes	yes	yes
F. construens	yes		yes	yes	yes
F. bicapitata	yes	yes	yes	yes	yes
F. intermedia			yes	yes	
F. zeilleri			yes	yes	
F. pinnata			yes	yes	yes
F. virscens			yes		
Gomphonema constritum				yes	
G. acuminatum		yes		yes	
G. intricatum		yes		yes	yes
G. subtile		yes		yes	
G. longiceps		yes		yes	
G. angrenensis		yes		yes	
G. olivaceum		yes			yes
Gyrosigma attenuatum	yes	yes		yes	yes

Diatom	**Diatom ecology**			**Water salinity**	
	Planktonic	**Epiphytic**	**Benthic**	**Fresh**	**Brackish**
G. scalproides	yes			yes	
Hantzschia amphioxys				yes	yes
Melosira granulata				yes	
Mastogloia elliptica		yes			yes
M. smithii		yes	yes	yes	yes
M. streptoraphe		yes		yes	yes
Nitschia denticula				yes	yes
N. frustulum				yes	yes
N. sinuate				yes	
N. angustata				yes	yes
N. hantzschiana				yes	
Navicula oblonga	yes	yes		yes	yes
N. radiosa		yes	yes	yes	yes
N. notha			yes	yes	
N. stroemii			yes	yes	
N. exilis			yes	yes	yes
N. schonfeldii			yes	yes	
N. diluviana			yes	yes	
N. tuscula			yes	yes	
N. pupula			yes	yes	yes
N. mutica			yes	yes	yes
N. rhynchocephala			yes	yes	yes
N. scutelloides			yes	yes	yes
N. lanceolata			yes	yes	yes
N. caroliniana			yes	yes	
N. gracilis			yes	yes	yes
N. cincta					yes
N. cryptocephala					yes
Neidium bisulcatum			yes	yes	
Opephora vulgara			yes	yes	
Pinnularia borealis		yes		yes	
P. cuneata		yes		yes	
Rhopalodia gibba	yes	yes		yes	yes
Rhoicosphenia curvata					yes
R.parallela	yes	yes			
R. currata	yes	yes			
Stauroneis acuta		yes	yes	yes	yes
S. phoenicenteron		yes	yes	yes	
Stephanodiscus astraea	yes				yes
S. rotula	yes				
Synedra capitata			yes	yes	
S. ulna	yes			yes	
S. intermedia				yes	
S. radians	yes		yes		yes
S. amphicephala				yes	
Tabellaria flocculosa	yes				
T. fenestrata	yes				

Table 4. Water depth and salinity indicated by ostracodes

Ostracode	Water depth		Water salinity				References
	Deep	Shallow	Fresh (<1 g/L)	Brackish (1-35 g/L)	Saline (> 35 g/L)	Measured salinity range (g/L)	
Candoniella leatea				yes	yes		Han and Dong, 1990
C. allicans		yes					Li YF et al., 1991
C. mirabilis		yes					Zheng et al., 1989
Candona neglecta	yes		yes				Huang et al., 1985
C. xizangensis	yes		yes				Huang et al., 1985
C. candida	yes		yes			< 2	Huang et al., 1985
C. rausoni	yes		yes				Huang et al., 1985
C. gyirongensis	yes		yes				Li YF et al., 1991
Cyprideis torosa			yes	yes	yes	<1 to 120	Li YF et al., 1991; 1997
C. littoralis			yes			2-5	Li YF et al., 1997
Cyprinotus spp.					yes		Zheng et al., 1989
Cundimella mirabilis				yes		20-30	Zheng et al., 1989
Dolerocypris fasciata			yes				Li YF et al., 1997
Eucypris gyirongensis			yes	yes		<5	Huang et al., 1985
E. inflata				yes	yes	8.5-257	Li YF et al., 1997
E. pigna			yes	yes			Huang et al., 1985
E. imglata					yes		Han et al., 1993
Ilyocypris biplicata		yes	yes	yes		0.51-280	Huang et al., 1985
I. bradi		yes	yes				Li BY et al., 1991
I. gibba		yes	yes			1.78	Li BY et al., 1991
I. fadyi			yes				Wu, 1995
Leucocythere mirabilis	yes	yes	yes	yes	yes	0.487-257	Li YF et al., 1997
L. postilirata				yes		8-10	Zheng et al., 1989
L. bispinosa			yes			> 1	Zheng et al., 1989; Li YF et al., 1994
L. binoda				yes			Zheng et al., 1989
L. dorsotuberosa			yes	yes		2	Wu, 1995
Leucocytherella sinensis			yes	yes		0-13	Huang et al., 1985
Limnocythere dubiosa			yes	yes		1-34	Li YF et al., 1997
L. sancti-patricii				yes			Zheng et al., 1989
L. inopinata	yes	yes	yes	yes		0-25	Wang SM et al., 1990
Limnocytherellina binoda					yes	80	Zheng et al., 1989
L. bispinosa			yes	yes	yes	0-173	Li YF et al., 1994
L. trispinosa			yes	yes		20	Zheng et al., 1989
L. kunlunensis			yes	yes			Zheng et al., 1989
Cypridopsis ofesa		yes	yes				Wu, 1995
Potamocypris			yes				Wu, 1995

Table 5. Water depth and salinity indicated by algae (after Institute of Hydrobiology, 1980)

Algae	Water depth			Water salinity
	Shallow	Intermediate	Deep	
Mougeotia	yes			fresh
Zygnema	yes			fresh
Spirogyra	yes			fresh
Pediastrum spp.			yes	fresh, brackish
Pediastrum boryanum			yes	fresh, brackish
Pediastrum simplex		yes		fresh, brackish

1.4.3. Standardisation: Lake Status Coding

For each site, an assessment of the relative water depth through time is made on the basis of all the available evidence. Where the evidence is limited or the range of variation is small, it may only be possible to distinguish two depth classes (deeper and shallower). There are sites, however, where it is possible to distinguish a greater number of depth-related differences in the geological and biostratigraphic data. This information is preserved by using a depth class categorisation that expands to the range demanded by the data. For each site, hiatuses are recorded by 0, the lowest water depth recorded is coded as status class 1 and then successively deeper water phases are coded as 2, 3, 4 and so on until the maximum water depth recognised in the basin has been coded. It should be noted that it is rarely possible to quantify the changes in depth and the status classes do not represent a linear scale of depth changes. Assessments of lake status, using this flexible coding scheme, are made on a continuous basis. This information, along with the specific basis for the depth categorisation, is given in the documentation file for each basin.

For comparison between sites, the lake status categories must be standardised. We use a 3-category scheme, where the status classes are defined as low (1), intermediate (2) and high (3). Various conventions can be used to define the boundaries of status classes. Here the boundaries were set so that for each lake record, the class "high" corresponded approximately to the upper quartile and "low" to the lower quartile of that lake's variation in level during the entire period of record, in order to ensure compatibility with the OLLDB (Street-Perrott et al., 1989). Status codings using this 3-category or "collapsed" status coding scheme have been made at 500 yr intervals for each lake. These collapsed status codings are given in the data base and have been used for e.g. mapping and subsequent analysis. It should be stressed, however, that alternative “collapsed” coding schemes are possible and can easily be derived from the continuous multi-class codings preserved in the data base. Similarly, data base users can extract information for any time interval(s) required. The 500 yr codings are presented here only to illustrate the nature of the data base.

1.4.4. Chronology and Dating Control

The chronology of changes in lake status at individual sites is generally based on radiocarbon dating (both conventional and AMS), thermoluminescence (TL) dating, Uranium-series dating, palaeomagnetic chronology, and other methods available from the literature. The chronology for a few sites is based on pollen-correlation with nearby radiometrically-dated sites or with a regional pollen chronostratigraphy. The top of a core or profile at a site where the sedimentation was known to be modern and continuous was used as if it were a date. Ages for stratigraphic boundaries were calculated by using linear interpolation between the nearest available dates.

The chronology of changes in lake status at each site is expressed on the radiocarbon time-scale (^{14} yr B.P.). We have made no attempt to calibrate the radiocarbon dates and correct the record to calendar years. There are two reasons for this decision. Firstly, many of the Chinese lake records are very long, and extend back beyond the current limit of calibration (20,265 yr B.P.: CALIB 4.3) (Stuiver and Reimer, 1993). Secondly, the existing records in the GLSDB are expressed in radiocarbon years. Thus our decision ensures compatibilty with existing regional datasets within the GLSDB. However, since all of the available information on the radiometric dates (e.g. date, error bar, depth, material dated) is preserved in the data base, as is the depth of lake status unit boundaries, database users may produce calibrated chronologies for individual sites relatively easily if required.

The quality of the dating varies between different parts of the record from a single basin, and between the records from different basins. The quality of the dating control has been assessed at each 500 yr interval, using ranking schemes developed for the Cooperative Holocene Mapping (COHMAP) project (Webb, 1985) and described by Yu and Harrison (1995). For data from continuous records, each interval was assigned a ranking (from 1 to 7) as follows:

1: Bracketing dates within a 2000 yr interval about the time being assessed
2: Bracketing dates, one within 2000 yr and the second within 4000 yr of the time being assessed
3: Bracketing dates within a 4000 yr interval about the time being assessed
4: Bracketing dates, one being within 4000 yr and the second being within 6000 yr of the time being assessed
5: Bracketing dates within a 6000 yr interval about the time being assessed
6: Bracketing dates, one within 6000 yr and the second within 8000 yr of the time being assessed
7: Poorly dated

This scheme is only appropriate when sedimentation is continuous so that interpolation between radiometric (or biostratigraphic) dates is possible. At some sites, the evidence for changes in water depth was derived from discontinuous sources, such as shorelines, lake deposits above the modern lake and archaeological features. In some lake cores there is evidence of reworking, slumping, sedimentary hiatuses or marked variations in sedimentation rates, all of which make it difficult to interpolate between the dated intervals. Even when sediment deposition can be assumed to be both continuous and uniform, there may be situations where there is a date very near one of the time intervals being coded but where the bracketing date is rather distant. Application of the scheme for continuous records would result in such an interval being inappropriately downweighted. For data from these types of discontinuous records (D), each 500 yr interval was assigned a ranking (from 1 to 7) as follows:

1: Date within 250 yr of the time being assessed
2: Date within 500 yr of the time being assessed
3: Date within 750 yr of the time being assessed
4: Date within 1000 yr of the time being assessed
5: Date within 1500 yr of the time being assessed
6: Date within 2000 yr of the time being assessed
7: Poorly dated

1.5. Structure of this Report

In this report, we introduce the general information for the Chinese Lake Status Data Base (Section 1) and describe the basic structure of the data base (Section 2). The documentation files describing the primary data and the reconstructed changes in lake status for every site are given in Section 3. A list of references on which the reconstruction is based is given at the end of each documentation file; a complete bibliography of the literature covering all the sites is given in Section 4. The site and dating information and the status codings at 500 yr intervals are summarised in a series of data base files. These files are listed in Appendix A. Finally, illustrative maps, showing the reconstructed patterns of lake status changes at 1000 yr intervals back to 18,000 yr B.P. are given in Appendix B.

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The photos on the cover were taken by Xia Weilan (NIGLAS). They show four lakes from the Tibetan Plateau: an unnamed lake in central Tibet (top left), Cuoer Lake (top right), Namucuo Lake (lower left) and Gounongcuo Lake (lower right).

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1.7. References to Introduction

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2. The Structure of the Data Base

The data base consists of documentation files (CHLAKE.DOC and CHREFS.DOC) and summary tables (CHDATA.XLS, CHSTATUS.XLS, CHCOLSTA.XLS, CHDC.XLS and CHDATLST.XLS). In the documentation files (Section 3) the sites are firstly arranged by alphabetical order of the region in which they occur, and within regions by alphabetical order of the basins. In the summary tables (Appendix A), the sites are arranged solely by the alphabetical order of the basins.

2.1. Content and Format of the Documentation Files

2.1.1. CHLAKE.DOC

These files are Word 98 files.

There is a separate documentation file for each site in the data base. The file contains a short description of the basin, including information (where available) on basin morphology, modern lake hydrology and catchment geology. The file also includes the reconstructed changes in lake status through time, along with a summary of the primary data on which these reconstructions are based. The basis for the status coding at each site is explicitly stated. The radiometric dates used to establish the chronology and the sources of data are listed. The status coding through time is also given. Most of the files are accompanied by simplified lithological diagrams of key cores or sequences used in the reconstruction.

2.1.2. CHREFS.DOC

This is a Word 98 file, containing full citation of all references used to compile the CLSDB.1. The references are listed in alphabetical order of author(s).

2.2. Content and Format of the Data Base Summary Files

2.2.1. CHDATA.XLS

This is an Excel 98 file containing the folowing primary information:

Basin name, province/region, latitude (in decimal degrees, N=+), longitude (in decimal degrees, E=+), elevation, water type, basin type, origin, basin geology, basin area, lake area, mire area, mean depth of lake water, maximum depth of lake water, basin precipitation, basin evaporation, number of ^{14}C dates, number of U-series dates, number of TL dates, number of palaeomagnetic dates, number of ^{210}Pb dates, number of other dates, record length, data sources (including lithology, hiatus, sedimentation rate, aquatic plants, diatoms, molluscs, ostracodes, geochemistry, geomorphology, archaeology, stable isotopes of ^{18}O and/or ^{13}C, others), primary references, other references, the persons responsible for the coding, and the date of final coding.

The data format is:

Basin, province/region, latitude (in decimal degrees, N=+), longitude (in decimal degrees, E=+), elevation (m), type of basin (overflow or closed lake, mire, bog etc.), basin origin and geology, area of basin (km^2), lake and mire area (km^2), mean and maximum depth of lake (m), number of ^{14}C, U-series, TL, palaeomagnetic, ^{210}Pb dates, record length (yr B.P.), data sources, primary references, other references, coded by (the people specifically responsible for the coding), date of final coding and comments.

The following abbreviations are used to indicate who was responsible for coding specific basins:

GY Ge Yu (Nanjing Institute of Geography & Limnology, Chinese Academy of Sciences, Nanjing 210008, China)

BX Bin Xue (Nanjing Institute of Geography & Limnology, Chinese Academy of Sciences, Nanjing 210008, China)

SPH Sandy P. Harrison (Max Planck Institute for Biogeochemistry, 07743 Jena, Germany)

2.2.2. CHSTATUS.XLS

This is an Excel 98 file containing:
Basin name, latitude, longitude, and the original status codings at 500-yr intervals from modern (0) to 30,000 yr B.P.

The data format is:
Basin name, latitude (in decimal degrees, N=+), longitude (in decimal degrees, E=+), status coding from 0ka back to 30 ka at 500 yr intervals (0 = hiatus, 1 = lowest, 2 = next lowest etc.; times that cannot be coded are marked n/c, times when alternative codings are possible because of dating uncertainties are shown with a slash, e.g. 2/1/0).

2.2.3. CHCOLSTA.XLS

This is an Excel 98 file containing:
Basin name, latitude, longitude, and collapsed status codings at 500-yr intervals from modern (0) to 30,000 yr B.P.

The data format is:
Basin name, latitude (in decimal degrees, N=+), longitude (in decimal degrees, E=+), collapsed status codings (1 = low, 2 = intermediate, 3 = high, n/c= not coded) at 500 yr intervals from 0-30 ka (see Section 1.4.3).

2.2.4. CHDC.XLS

This is an Excel 98 file containing:
Basin name, latitude, longitude, and dating control.

The data format is:
Basin name, latitude (in decimal degrees, N=+), longitude (in decimal degrees, E=+), dating control at 500 yr intervals from 0-30 ka (code see Section 1.4.4).

2.2.5. CHDATLST.XLS

This is an Excel 98 file containing:
Basin name, latitude, longitude, number of dates, laboratory number for each date (when available), date, positive error and negative error, sample depth, description of material,

core/profile name, and notes. A distinction is made in the Note column between AMS ^{14}C dates, U/Th and TL dates; all other dates are conventional ^{14}C dates.

The data format is:
Basin name, country, latitude (in decimal degrees, N=+), longitude (in decimal degrees, E=+), total number of dates, laboratory number, date (yr B.P.), positive error and negative error, sample depth of date (m), description of material, name of core or profile if there are multiple cores or profiles from the basin, comments (ATO = age too old, ATY = age too young possible, AMS = AMS dating, TL = TL dating, U/Th = U/Th dating). The comments section includes further details of the reasons why specific dates are not used to erect a site chronology.

3. Lake Status Records from China

In this section, we document the records of 42 lakes from China. The sites are arranged in alphabetical order of the regions in which they occur, and within regions by alphabetical order of the basins. The regions are based on the political divisions of China (Fig. 4).

Figure 4: The political regions of China

3.1. Baijian Lake, Gansu Province

Baijian lake (39°09'N, 104°10'E, ca 1282 m a.s.l.) was a closed palaeolake, located in modern Tenggar desert (Pachur et al., 1995; Zhang and Wunnemann, 1995). The basin currently holds a saline swamp, with brackish water about 1 m below the surface. The catchment area is > 91,000 km². The lake basin is of tectonic origin. The evidence of neotectonic movement is distinct in the southern part of the basin, but weak in the north. Tectonism is not thought to have influenced the lake record during the Late Quaternary. The basin is bounded by the Qilian Mountains in the south, the Helan Mountains in the eastt, and Yabulai Mountains and the Gobi Plateau in the west and the north. The bedrock consists of Precambrian and Palaeozoic metamorphic and crystalline rocks, and Mesozoic, Tertiary and Quaternary limestones and sandstones. There are only three permanent rivers flowing into the lake: Shiyang River, Hei River and Shure River, all of which originate from Qilian Mountain. The lake is also fed by some intermittent rivers which flow during the summer season. Qilian Mountain is glaciated above elevations of ca 4900-5500 a.s.l. The regional climate is under the control of the East Asian monsoon. The annual precipitation is ca 115 mm, and the potential evaporation is ca 2600 mm.

The existence of a large palaeolake within the Baijan bain is indicated by lacustrine terraces at elevations from +4 to +31m above the level of the modern swamp. A more detailed record of relative water depth changes is provided by the lithological record from 3 cores (B100, Duantouliang, Tudungcao) from the basin floor and two profiles (Jiajiakeng and Magang) from within the basin. The chronology is based on 15 radiocarbon dates from the terraces, 7 dates from core B100, 5 dates from the Duantouliang core, 3 dates from the Tudungcao core, 4 dates from the Jiajiakeng core, and 4 dates from the Magang core.

Six well-preserved lacustrine terraces, indicating high lake stands during the last 40,000 yr B.P., are found in the northern part of the basin (Pachur et al., 1995; Zhang and Wunnemann, 1995). These terraces appear to be constructional features, containing mollusc shells and beach materials, and are built on sandy lake carbonate deposits. The following table summarises the information available about the 6 terraces:

Terrace Name	Elevation (m a.s.l.)	Height above lake (m)	^{14}C dates	Source/Comments
T1	1312	30	undated	Ref 1
above T2.1	1309	27	33,500±1085; 32,435±840	Ref 1; T2.1 in Ref 2
above T2.1	1309	27	32,270±1236	Ref 1; T2.2 in Ref 2
T2.1	1306	24	30,330±560; 27,200±975; 23,370±380	Ref 1; T2.3 in Ref 2
T2.2	1304	22	23,130±590; 16,540±120; 12,817±140	Ref 1
T2.3	1280.2		22,886±180	Ref 2; but text indicates date from 1.8m pit in modern lake floor
T3	1296	14	undated	Ref 1
T4	1295	13	5250±70; 5510±60	Ref 1; Ref 2
T5	1290	8	3660±55	Ref 1; Ref 2
T6	1286	4	1910±60; 1405±60	Ref 1; Ref 2

Ref 1: Pachur et al., 1995; Ref 2: Zhang and Wunnemann, 1995

The nomenclature of the terraces differs between the various publications, and the descriptions of these features and of the locations from which the radiocarbon-dated material was taken is very imprecise. It is only possible to reconcile the authors' interpretations of lake-level changes within the basin by assuming that the older dates on some of the terraces (e.g. Terrace T2.2) comes from the lacustrine sandy carbonate deposits that underlie the constructional terraces themselves. However, even making this assumption, it is not possible to reconcile all of the available data from the terrace sequence with the lake-level reconstructions described by the original authors. Here, we have tried to derive a consistent interpretation of the available data. Discrepancies between our interpretations and those of the original authors are identified.

The terraces are underlain by sandy lake carbonates, characteristic of the littoral zone. Since these deposits extend to an elevation of ca 31m above modern lake level, the lake itself must have been extremely large. The three radiocarbon dates from "above T2.1" were likely taken from exposures of this sandy carbonate between the T1 and T2.1 beach terraces. These dates suggest that the sandy carbonate unit was formed ca 33,500 to 32,270 yr B.P. The highest terrace (T1), at an elevation of ca 1312 m a.s.l. (+30 m above the modern lake floor), is characterized by mollusc-bearing beach sand and gravel. The deposits have not been radiocarbon dated, but given their superposition over the sandy carbonate unit probably formed ca 32,000 yr B.P.

The second to the highest terrace (T2), at an elevation of ca 1304-1309 m a.s.l. (+22 to 27 m above the modern lake floor) and ca 1200m away from the margin of the lake floor, is the most distinct of the terraces. It is therefore thought to represent a relatively long-lived high lake stand (Pachur et al. 1995; Zhang and Wunnemann, 1995). It appears to consist of three separate ridges (T2.1, T2.2, T2.3). Radiocarbon dates on the outermost ridge (T2.1) suggest it was formed between 30,330 and 23,370 yr B.P. During this relatively stable interval of high lake levels (+24m above modern lake level) the lake would have had an area of ca 16200 km². There are three radiocarbon dates associated with the T2.2 terrace: 23,130±590 yr B.P., 16,540±120 yr B.P. and 12,817±140 yr B.P. These dates could be interpreted as suggesting that the T2.2 terrace ridge started forming immediately after formation of the T2.1 ridge and continued for ca 13,000 years. This seems unlikely, especially given evidence from the core sequences that there was a major drying phase occurring around 18,000 yr B.P. If we assume that the basal date from T2.2 (23,130 yr B.P.) dates the underlying sandy carbonate lacustrine sediments, then the record can be interpreted as showing continued lacustrine deposition at lower elevations during formation of T2.1 and terminating, according to both the dates on T2.1 and the basal date on T2.2 at ca 23,000 yr B.P. A return to wetter conditions is indicated by the formation of the T2.2 beach ridge. The lake level would have been ca +22m above modern lake level. The radiocarbon dates suggest this occurred between ca 16,540 and 12,817 yr B.P. The intervening interval must then have been characterised by water levels lower than the T2.2 beach ridge (i.e. below 1304m). Pachur et al. (1995) suggest that the T2.2 beach ridge was only formed after ca 13,000 yr B.P., but provide no explanation for the older date from this feature.

The third terrace (T3), at an altitude of 1296 m a.s.l., consists of a beach bar containing abundant molluscs. This terrace has not been radiocarbon dated, but is presumably younger than T2.2 (i.e. younger than 12,000 yr B.P.).

The T4 terrace, ca 1295 m a.s.l., is also characterized by beach sand and gravel. Two samples of snail shell and carbonate from this terrace have been radiocarbon dated to

5250±70 and 5510±60 yr B.P. respectively. These dates suggest the lake level was +10m above modern lake level between 5600 and 5000 yr B.P.

The T5 terrace lies at an altitude of ca 1290 m a.s.l. (+8m above the elevation of the modern lake floor). A single snail shell sample has been radiocarbon dated to 3660±55 yr B.P., suggesting the lake was relatively high ca 3600-3700 yr B.P.

The T6 terrace lies at an altitude of 1286 m a.s.l. (+4m above the elevation of the modern lake floor). The terrace consists of lacustrine carbonate deposits. There are no radiocarbon dates from these deposits. However, two samples on the carbonate from a nearby section at the same altitude have been radiocarbon dated to 1910±60 and 1405±60 yr B.P. This suggests that the lake was ca +3m above its modern elevation ca 2000-1400 yr B.P.

Assuming that the lake was low during the intervals between formation of these terraces, the record of lake-level changes during the last 33,500 yr appear to show an oscillatory pattern but with high stands at successively lower elevations. Thus, the lake was initially high during the deposition of the sandy carbonate deposits (up to 1312 m a.s.l.) before 33,500 yr B.P. The lake was again high (ca 1306-1308 m a.s.l.) between 30,330 and 23,370 yr B.P. Between 16,000 and 12,000 yr B.P. the lake rose again to ca 1304 m a.s.l. The remaining three terraces indicate high stands during the Holocene, although only one (T4) has been directly dated. This terrace indicates that the lake level was at 1295 m a.s.l. between 5500-5000 yr B.P. The youngest terrace in the basin (at ca +3m above modern lake floor elevation) is thought to have formed shortly after 2000 yr B.P.

There are three cores and two profiles in this basin. A 70 m core (B100) was taken from the center of the modern lake floor at ca 1280 m a.s.l. The coring site probably represents the center of the Baijian palaeolake, and thus should provide a deep-water record of the lake. The core provides a sedimentary record covering the last 19,000 yr B.P. and the interval before ca 27000 yr B.P. The record between 27000-19000 is absent due to lack of core material. Two other cores (the 3.8 m-long Duantouliang core and the 3.3 m-long Tudungcao core), taken from the northwestern part of the basin at an elevation of ca 1266 m a.s.l. (Ma et al., unpublished manuscript), provide a sedimentary record for the interval ca 39,000-23,000 yr B.P. Two 2.6 m long profiles (the Jiajiakeng and Magang profiles), taken from the western part of the basin at an altitude of ca 1295 m a.s.l., provide a sedimentary record of the interval between ca 12,000-5800 yr B.P.

There are no radiocarbon dates below ca 11m in core B100. The unit between 13.8-11 m is aeolian sand, suggesting the lake basin was dry. The overlying unit (11-7.8 m) is clayey silt of lacustrine origin, suggesting the onset of lacustrine conditions. Two samples from 9.3 m and 8.2 m are radiocarbon dated to 35,660±420 yr B.P. (AMS) and 27,150±615 yr B.P. respectively. There is another AMS radiocarbon date of 31,060±220 yr B.P. from a depth of 7.0 m in core B100. Using the sedimentary rate between the two AMS dates (0.0129 cm/yr), the lacustrine phase indicated by the clayey silt unit occurred between ca 39,000 and 27,000 yr B.P. The overlying unit (7.8-6.0 m) is fluvial gravel and sand, interbedded with a very thin layer of sandy silt. The presence of fluvial deposits in the central part of the basin indicates that lake levels had dropped considerably around ca 27,000 yr B.P. The nature of the unit between 6.0-3.2 m is not described in the published papers. The overlying unit (3-2.5 m) is mollusc-bearing clayey silt, suggesting that the lake was relatively deep, although shallower than in the interval between 39,000-27,000 yr B.P. A sample from 2.9 m is radiocarbon dated to 18,620±325 yr B.P., suggesting this phase occurred before ca 18000 yr B.P. (Pachur et al., 1995). The overlying unit (2.5-1.15 m) is well-sorted, fine to medium aeolian sand, suggesting the lake basin became dry after 18,000 yr B.P. The overlying unit (1.15-1.0 m) is sandy carbonate of lacustrine origin, suggesting a return to lacustrine conditions. The sandy, carbonate-rich nature of the deposits is consistent with relatively shallow-water lacustrine conditions. By extrapolation of the sedimentation rate (0.00988 cm/yr) on the overlying units, this phase occurred between ca 9200-7600 yr B.P. The overlying unit (1.0-0.25 m) is sandy carbonate alternating with chalk. The increase in the abundance of sodium chloride and gypsum suggests the lake became shallower. Two samples from within this unit (0.9 m and 0.57 m) are radiocarbon dated to 6655±100 and 3315±130 yr B.P. respectively. Extrapolation of the sedimentation rate between these two dates suggests this shallower-water phase occurred between ca 7600-1600 yr B.P. The uppermost unit (0.0-0.25m) is salty clay. The high salt content indicates further shallowing after 1600 yr B.P.

It is difficult to correlate the Core B100 record with the record shown by the terraces. The basal lacustrine unit in Core B100 (39,000-27,000 yr B.P.) probably correlates with the sandy carbonate littoral-zone deposits underlying the terraces (33,500-32,270 yr B.P.) and the formation of the T1 terrace. The phase of lacustrine deposition between 30,000 and 23,000 yr B.P. implied by the formation of the T1 terrace and the 23,000 yr B.P. date on deposits at the base of the T2.2 terrace could be correlated with the lacustrine phase recorded by the B100 core as occurring prior to ca 18,000 yr B.P. However, there is no evidence of a lacustrine phase occurring ca 16,540 to 12,800 yr B.P. (as evidenced by the T2.2 terrace) in the core sediments. However, the sequence of lacustrine sediments which show three phases of progressive shallowing after ca 9200 yr B.P. could be correlated with the undated T3 terrace, the T4 and T5 terraces, and the T6 terrace respectively.

The basal unit (3.8-3.53 m) in the Duantouliang core is yellow brown gravel, suggesting shallow water condition in the lake basin. This phase occurred before 42,000 yr B.P. (Ma et al., unpublished manuscript). The overlying unit (3.53-2.2 m) is pale grey lacustrine clay interbedded with carbonate-rich lacustrine silty clay. The nature of these deposits indicates relatively deep-water lacustrine conditions in the basin. The presence of the freshwater ostracode *Limnocythere inopinata* is consistent with relatively deep water (Pachur et al., 1995). Two samples from 3.1-3.2 m and 2.75-2.7 m in the Duantouliang core are radiocarbon dated to 38,650±970 and 35,020±810 yr B.P. respectively, suggesting this phase of deep-water lacustrine conditions occurred between ca 42,000 and 31,000 yr B.P. The overlying unit (2.2-1.35 m) is a mollusc-bearing, pale grey clay interbedded with carbonate-rich silty clay. Although there is no apparent change in the lithology, the presence of molluscs suggests somewhat shallower conditions than before. This shallower phase occurred ca 31,000-23,410 yr B.P. The overlying unit (1.35-0.69 m) is a mollusc-bearing, sandy gravel. The deposits are consist with beach or nearshore deposits, suggesting a further shallowing after 23,410 yr B.P. The overlying unit (0.69-0.25 m) is carbonate-rich clayey silt. The change in lithology indicates that the lake became deeper sometime after ca 17,820 yr B.P. The uppermost unit (0.25-0 m) is a mollusc-bearing, sandy gravel bach deposit, suggesting the lake became shallower again. By extrapolation of the sedimentation rates between the radiocarbon dates from this core, this interval of shallowing occurred ca 14,000 yr B.P.

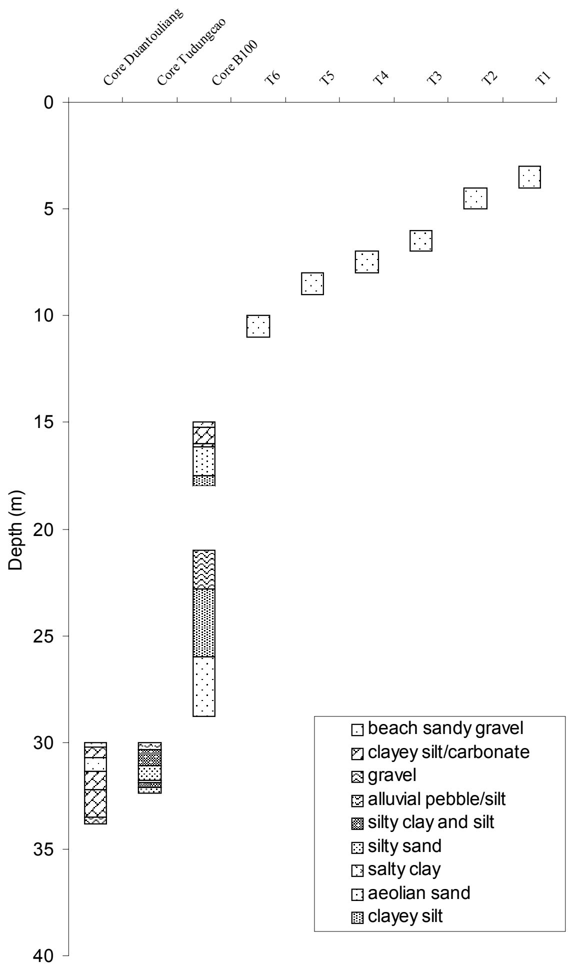
It is difficult to reconcile the lithological changes recorded in the Duantoulong core with its low-elevation position in the basin, without postulating substantial geomorphological changes. Nevertheless, the sequence and timing of the implied changes in water depth shows a remarkable consistency with the terrace records. Thus, the initial deepwater phase (42,000-31,000 yr B.P.) can be correlated with the initial deepwater phase implied by the sandy carbonate deposits that underlie the highest terraces (35,000-32,000 yr B.P.). The subsequent phase of shallower conditions (31,000-23,400 yr B.P.) correlates with the formation of the T2.1 terrace. The subsequent phase of shallower conditions (23,400-17,800 yr B.P.) can be correlated to the lake recession that occurred between formation of the T2.1 and T2.2 terraces. The return to deeper conditions (17,800-14,000 yr B.P.) can be correlated with the formation of the T2.2 terrace (16,000-12,800 yr B.P.) and the final phase of shallowing can be correlated with the decrease in lake level that occurred after the formation of the T2.2 terrace.

There are no radiocarbon dates below ca 2.1m in the Tudungcao core. The unit between 2.4-2.1 m is a mollusc-bearing silty sand. The lithology is consistent with shallow-water deposition. The overlying unit (ca 2.1-1.9 m) is lacustrine silty clay, suggesting the lake became deeper. A sample from ca 2 m is radiocarbon dated to 38,860±920 yr B.P., suggesting this deeper-water phase occurred ca 40,000-38,000 yr B.P. The overlying unit (1.9-1.8 m) is a mollusc-bearing sandy gravel beach deposit, suggesting shallowing ca 38,000-37,000 yr B.P. The overlying unit (1.8-1.1 m) is a mollusc-bearing silty sand. The lithology is consistent with nearshore deposition, suggesting that the lake became somewhat deeper than before. A sample from 1.69-1.79 m is radiocarbon dated to 36,625±1630 yr B.P., suggesting this somewhat deeper-water phase occurred ca 37,000-26,000 yr B.P. The overlying unit (1.1-0.35 m) is silty clay and silt which suggest a further deepening of the lake. A sample from ca 1.05 m is radiocarbon dated to 25,920±900 yr B.P. Interpolation between this date and the underlying radiocarbon dates suggests that this deeper water phase occurred ca 26000-19000 yr B.P. The overlying unit (0.0-0.35m) consists of pebbles and silt, and is of alluvial origin. This lithological change indicates lake level dropped dramatically after 19,000 yr B.P.

It is not possible to correlate the record from Tudungcao with the record from Duantuoliang, despite the fact that both cores occur in the same part of the basin and apparently at the same elevation. Given the relatively few radiocarbon dates available, and the fact that these are dates close to the limits of radiocarbon reliability on bulk samples, we assume that this core bears witness to oscillations within the earliest phase of deepwater conditions within the basin.

The 3.7m deep profiles at Jiajiakeng and Magang (1295 m a.s.l.) are characterized by silt and thinly-bedded clayey silt. The lithology suggests relatively deep water conditions. The 6 radiocarbon dates from these two sections suggest that this phase occurred ca 13000-5800 yr B.P. Beach shells have been found on the surface close to the site of these profiles. One sample of the shells has been radiocarbon dated to 8720±105 yr B.P., consistent with the idea that the lake was relatively high during the early Holocene. The elevation of the coring sites is the same as the T4 and within 1m of the elevation of the T3 terrace. The deposits appear to be too old to be correlated with the formation of the T4 terrace (5600-5000 yr B.P.) but could be correlated with the T3 terrace. In this case, the Jiajiakeng and Magang records suggest that the T3 terrace (which is undated) formed sometime in the interval between ca 13,000 to 5800 yr B.P. This is consistent with its position relative to the dated T2.2 and T4 terraces

In the status coding, extremely low (1) is indicated by modern conditions, and by alluvial and/or aeolian deposition (indicating phases when the coring site was dry) in Core B100; very low (2) by the T6 terrace (1286 m a.s.l.); low (3) by the T5 terrace (1290 m a.s.l.); moderately low (4) by the T4 terrace (1295 m a.s.l.); intermediate (5) by the T3 terrace (1296 m a.s.l.) and thinly-bedded clayey silt in the Jiajiakeng and Magang sections at 1295 m a.s.l.; moderately high (6) by sandy gravels in the Duantouliang core corresponding to shallow conditions between the T2.1 and T2.2 terrace formation (and therefore lower than the T2.2 terrace); high (7) the T2.2 terrace (1304 m a.s.l.) and carbonaceous clayey-silt deposition in the Duantouliang core; very high (8) by the T2.1 terrace and mollusc-bearing grey lacustrine clays in the Duantouliang core; extremely high (9) by sandy carbonate deposits underlying the T1 terrace, pale grey lacustrine clays in the Duantouliang core, and lacustrine silts in core B100.



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Radiocarbon Dates

Hv 18934	38860±920	ca 2 m	shells, Tudungcao
	38650±970	3.2-3.1 m	carbonate, Duantouliang
Lu 9310/N102	36625±1630	1.69-1.79 m	shells, Tudungcao
	35660±420	9.3 m	AMS, core B100
Lu 9324/N112	35020±810	2.75-2.7 m	marl, Duantouliang
Lu 934/N14	33500±1085		shells, above T2.1
Lu 9323/N111	33265±800		marl, Duantouliang
Lu 9315/N123	32435±840		shells, above T2.1
Lu 935/N18	32270±1236		shells, above T2.1
	31060±220	7.0 m	AMS, core B100
Ld 9310/N115	30360±175		shells, Duantouliang
Lu 936/N19	30330±560		shells, T2.1
Lanzhou Univ.	27200±975		shells, T2.1
Hv 19982	27150±615	8.2 m	carbonate, core B100; ?ATY, not used
Hv 19981	26900+1055/-890	6.7 m	carbonate, core B100
Ld 9311/N114	26749±164	1.55-1.5 m	carbonate, Duantouliang
Lanzhou univ.	25920±900	ca 1.05 m	shells, Tudungcao
Hv 19664	23370±380		shells, T2.1
Lanzhou Univ.	23130±590		carbonate, T2.2 (main beach)
Lu 9305/N221	22886±180		carbonate, from 1.8m deep pit on playa
Hv 19980	18620±325	2.9 m	carbonate, core B100
Hv 18936	16540±120		carbonate, T2.2 (main beach)
Lu/N 153	12817±140		carbonate, T2.2 (main beach)
Lu 9321/N42	12235±90		lacustrine clay, Jiajiakeng profile
Lu 9319/N41	12185±90		lacustrine clay, Jiajiakeng profile
Lu 9320/N43	10875±70		lacustrine clay, Jiajiakeng profile
Lu 931	8720±105		shells in sand-gravel beach, near Jiajiakeng
Lu 937/N21	8565±140		molluscs, Magang
Ld 9301/N32	8211±115		organic carbon, Magang
Hv 18933	7285±100		carbonate, Magang
Hv 19979	6655±100	0.9 m	carbonate, core B100
Lu 938/N31	5825±160		snails, Magang
Lu 9322/N211	5510±60		carbonate T4
Lu 933/N16	5250±70		snails, T4
Ld 9302/N91	4645±120		peat, Baguamiaou
Lu 932/N17	3660±55		snails, T5
Hv 19978	3315±130	0.57 m	carbonate, core B100
Ld 9303/N92	2561±85		peat, Baguamiaou
Lu 9318/N20	1910±60	bottom of section	carbonate, T6
Hv 18937	1405±60	top of section	carbonate, T6

Coding

39,000-32,000 yr B.P.	extremely high (9)
32,000-23,000 yr B.P.	very high (8)
23,300-19,000 yr B.P.	extremely low (1)
19,000-18,000 yr B.P.	very high (8)
18,000-16,000 yr B.P.	extremely low (1)
16,000-12,800 yr B.P.	high (7)
13,000-5800 yr B.P.	intermediate (5)
12,800-9200 yr B.P.	extremely low (1)
9200-7600 yr B.P.	intermediate (5)
7600-5600 yr B.P.	extremely low (1)
6600-3700 yr B.P.	moderately low (4)
5000-3500 yr B.P.	low (3)
3600-1400 yr B.P.	very low (2)
1600-0 yr B.P.	extremely low (1)

Preliminary coding: 01-12-1998
Second coding: 30-07-2000
Third coding: 06-09-2000
Final coding: 23-01-2001
Coded by BX, GY and SPH

3.2. Nancun, Guangxi Province

Nancun (24°45'N, 110°25'E, ca 160 m a.s.l.) was a small lake, now artificially drained for farmland. On the basis of the extent of peat deposits in the basin, the lake area before drainage was estimated to have been approximately 2.5 km² (Yao and Liang, 1993). There was no surface inflow and the catchment area of the former lake was confined to the small hills (ca 200 m a.s.l.) surrounding the lake. The annual mean precipitation is ca 1560-2060 mm, and over 60% of the precipitation occurs in summer. The annual evaporation is ca 1255 mm. The basin bedrock is sandstone, shale and carbonate. On the basis of the pollen record (specifically increases in Gramineae and *Pinus*) and historical records, Yao and Liang (1993) suggest the basin has been significantly impacted by human activities since ca 2400 yr B.P.

A transect of five cores (Core 1, Core 4, Core 9, Core 6 and Core 5) taken across the former lake provides a stratigraphic record back to ca 6400 yr B.P. (Yao and Liang, 1993). None of the cores reach the bottom of the lacustrine deposits. Core 9 (2.45m) was taken from the central part of the transect and was analysed for lithology and pollen. There are four 4 radiocarbon dates from this core. The remaining cores were taken from nearer to the lake margin, and have not been analysed in detail. Two samples from Core 4 (ca 2.2m long) were radiocarbon-dated. A further two dates have been obtained on the top and bottom of the peat layer which occurs in the stratigraphic section, but there is no indication which core was dated. Changes in relative water depth are reconstructed on the basis of changes in lithology shown by the core transect and the aquatic pollen from core 9.

The basal unit (2.45-2.36 m in Core 9; 2.20-1.90m in Core 4) is greyish yellow lacustrine clay. The lithology indicates moderately deep water conditions. The low abundance of Cyperaceae (ca 250 $g/cm^2 \cdot a$ as shown as influx) in the pollen record from Core 9 is consistent with this interpretation. A sample from 2.45 m in Core 9 is radiocarbon dated to 5875±74 yr B.P. Linear interpolation between this date and the date at 2.1 m in Core 9 indicates this deep-water phase occurred ca 5880-5750 yr B.P. However, a basal sample from Core 4 was radiocarbon-dated to 5300±92 yr B.P. Linear interpolation between this date and a date at 1.6m, would suggest the upper boundary of lacustrine clay dates to ca 4540 yr B.P.

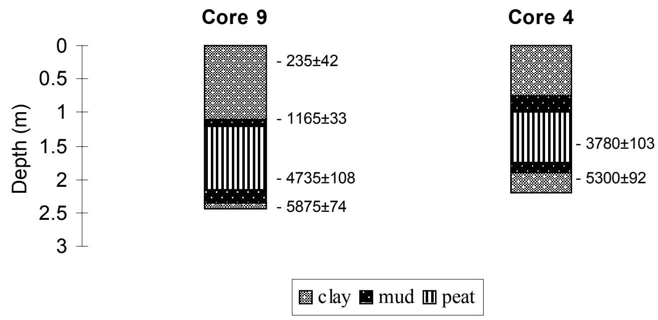
The overlying unit (2.36-2.15 m in Core 9; 1.90-1.75m in Core 4) is black mud. The increase in organic content suggests the lake became shallower. The abundance of Cyperaceae (ca 250 $g/cm^2 \cdot a$) in the pollen record from Core 9 increased slightly, consistent with shallowing. By interpolation, this phase as recorded in Core 9 occurred ca 5750-5020 yr B.P. The age estimates on Core 4 are ca 4540-4160 yr B.P.

A further shallowing is indicated by peat deposition (2.15-1.2 m in Core 9; 1.75-1.00m in Core 4). The coring survey indicates that the thickness of this peat unit decreases towards the lake margin. The abundance of Cyperaceae (ca 100-500 $g/cm^2 \cdot a$) in the pollen record from Core 9 increased, consistent with shallowing. A sample from 2.1 m in Core 9 was radiocarbon dated to 4735±108 yr B.P. By linear interpolation between this date and the date from the overlying unit, this phase occurred ca 5020-1350 yr B.P. Two samples from the depth of 2.2 m and 1.6 m in core 4 are radiocarbon dated to 5300±92 and 3780±103 yr B.P. respectively, indicating peat deposition (ca 1.75-1 m) occurred ca 4160-2260 yr B.P. However, two dates on the top and bottom of this peat layer from an unspecified core from the basin, indicate that peat deposition occurred between 6400 and 1630 yr B.P.

The overlying unit (1.2-1.1 m in Core 9; 1.00-0.75 m in Core 4) is black mud, suggesting the lake became deeper. The abundance of Cyperaceae (ca 200 $g/cm^2 \cdot a$) in the pollen record from Core 9 decreased, consistent with deepening. By interpolation, this phase occurred ca 1350-1120 yr B.P. according to interpolation in Core 9, while ca 2260-1626 yr B.P. according to interpolation in Core 4.

The overlying unit (1.1-0 m in Core 9; 0.75-0 m in Core 4) is greyish black brown clay. The abundance of Cyperaceae (ca 250 $g/cm^2 \cdot a$) in pollen record from Core 9 decreased. Two samples from 1.15 and 0.05 m in Core 9 were radiocarbon dated to 1165±33 and 235±42 yr B.P. respectively, suggesting this phase occurred ca 1200-0 yr B.P. The distinct increase in sedimentary rate to 0.96 mm/yr from ca 0.3 mm/yr in the underlying unit is consistent with enhanced human activities.

It is not possible to reconcile or explain the differences in chronology between the three radiocarbon-dated sequences. The minimum interpretation of the records suggest the lake was low (1) from ca 6400 to ca 1350 yr B.P., rose to intermediate levels (2) between 1350 and 1120 yr B.P. and was high (3) thereafter. The lake may have been impacted by human activities in the last 1000-2000 years, but this does not seem to have affected the lake level.



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Radiocarbon dates

GL-82066	6400±115	bottom of peat layer, core unspecified
ZD2-340	5875±74	2.45 m, clay, Core 9
ZD2-345	5300±92	2.2 m, clay, Core 4
ZD2-341	4735±108	2.1 m, peat Core 9
ZD2-344	3780±103	1.6 m, peat, Core 4
GL-82006	1630±100	top of peat layer, core unspecified
ZD2-342	1165±33	1.15 m, clay, Core 9
ZD2-343	235±42	0.05 m, clay, Core 9

Coding

6400-1350 yr B.P.	low (1)
1350-1120 yr B.P.	intermediate (2)
1120-0 yr B.P.	high (3)

Preliminary coding: 01-12-1998
Second coding: 01-01-1999
Final coding: 01-03-1999
Coded by BX, GY and SPH

3.3. Ningjingbo, Hebei Province

Ningjingbo (37°-37°30'N, 114°40'-115°15'E, ca 24-28 m a.s.l.) is a basin in the southern part of the Hebei Plain, and is surrounded by the Taihang Mountains. The basin is tectonic in origin and was formerly occupied by a shallow, freshwater lake that occurred in the lowland between the alluvial fans of the Zhang and Futuo Rivers. Several hundred years ago, the lake started to dry out as a result of climatic changes and human activities (Guo and Shi, 1999) such that it was completely dry by 1839 A.D. The basin is dry today except during flood seasons, when a shallow freshwater lake is temporarily formed. The former lake was fed by two rivers, the Zhang River from the south and the Futuo River from the north, and several small streams which arise in the Taihang Mountains. There was an outflow via the Fuyang River to the Bohai Sea, which still operates today in flood seasons. The Fuyang River cuts through the Taihang Mountains and flows ca 300 km across the Beijing Plain to the Bohai Sea. The regional climate is monsoonal, with an annual mean precipitation of ca 500 mm, most of which is concentrated in the summer season. The annual mean temperature is ca 12-13 ^{o}C and the summer temperature is 26.7^{o}C (Guo, 2000).

The stratigraphy of the lacustrine deposits at Ningjingbo have been studied from a 10m-deep outcropping section in the central part of the now-dry lake (Guo and Shi, 1999) and from a 42.8m-deep core (Nanwangzhuang Borehole) about 20m away from the section (Guo, 2000). The section and the Nanwangzhuang Borehole core stratigraphy is the same, although the uppermost meter of the Borehole deposits is anthropogenic and not seen in the section. Changes in relative water depth can be reconstructed from changes in lithology, aquatic pollen assemblages and the presence or absence of molluscs, as shown in both sequences (Guo, 2000, Guo and Shi, 1999), and from changes in ostracode assemblages as shown in the upper 6m of the Borehole core (Guo, 1998). The chronology is based on four radiocarbon dates from material in the exposed section (Guo, 2000, Guo and Shi, 1999). These dates have been transferred to the core by adding 1m to the stated depth (to allow for the anthropogenic material at the top of the core). A second core from the eastern part of the basin (Julu) has not been studied in detail, although a single radiocarbon date has been obtained from this core (Guo, 2000). On the basis of correlation with the regional pollen chronostratigraphy, the Nanwangzhuang Borehole core is thought to extend back beyond the last glacial maximum. However, all of the available radiocarbon dates are Holocene in age, and thus it is not possible to erect a reliable chronology for the earliest part of the record.

The basal unit (42.8-37.0 m) in the Nanwangzhuang Borehole core is greyish-green silt. The lithology and the presence of *Typha* suggests that this unit is lacustrine in origin.

The overlying unit (37.0-23.6 m) in the Nanwangzhuang Borehole core is brown to greyish-green clay, with horizontal bedding. The preservation of horizontal bedding suggests the lake became deeper. The absence of *Typha* from most samples within this unit is consistent with this interpretation.

The overlying unit (23.6-18.1 m) in the Nanwangzhuang Borehole core is brown clay and silt, with horizontal bedding. The increase in silt content suggests that the water depth decreased. The unit contains mollusc shells, consistent with shallower conditions. *Typha* is present in occasional samples.

The overlying unit (18.1-13.9 m) in the Nanwangzhuang Borehole core is greyish-brown to greyish-white poorly-sorted medium to coarse sand, containing fine gravel and calcite concretions. Mineralogical analysis shows the predominance of the less-stable minerals, indicating that this material was laid down quickly and not subjected to weathering. Thus the lithology and mineralogy suggest this unit is of alluvial origin (Guo, 2000). The deposition of alluvial fan deposits at the coring site indicates the basin was dry. The terrestrial pollen assemblage is characterised by the dominance of non-arboreal taxa, and the unit is thus thought to have been formed some time during the last glacial maximum (Guo, 2000, Guo and Shi, 1999).

The overlying unit (13.9-6.6 m) in the Nanwangzhuang Borehole is greyish-brown to greyish-green silty clay and silt, with horizontal bedding. This unit marks the return of lacustrine conditions in the basin. The aquatic pollen assemblage is characterised by a sequence in which *Typha* is dominant initially and is then replaced first by Alismataceae (0-20%) and subsequently by Potamogetonaceae. This is consistent with a gradual increase in water depth throughout the interval. According to Guo (2000), this phase occurred in the late glacial. A sample from 5.6m in the outcropping section, equivalent to 6.6m in the core, was radiocarbon dated to 9750±350 yr B.P., indicating that this return to lacustrine conditions occurred sometime before 9750 yr B.P.

The overlying unit (6.6-5.2 m) in the Nanwangzhuang Borehole is black to dark grey muddy clay, with horizontal bedding. The increase in clay content suggests the lake became deeper. The absence of Alismataceae, and the abundance of Potamogetonaceae (10-20%) in the aquatic pollen are consistent with increased water depth. The abundance of *Typha* may reflect the extension of the fringing wetlands. The ostracode assemblage is dominated by *Candona*, consistent with relatively deep conditions. Towards the top of the unit (5.4-5.2 m) there are distinct layers with abundant calcareous nodules. Between these layers, there is a ca 10cm thick layer with very abundant mollusc shells. Although the assemblage is characterised by very few species, it includes both young and old examples of each species in growth position and is therefore clearly in situ. The increased organic content, the presence of calcareous nodules and the abundance of molluscs is indicative of a significant shallowing. The ostracode assemblage in this uppermost part, is dominated by *Ilyocypris*, consistent with decreased water depth. This unit is recognised in the outcropping section, where samples from the top and bottom boundaries (4.2 and 5.6m) have been radiocarbon-dated to 9750±350 and 5277±157 yr B.P. respectively. This suggests that the initial phase of relatively deep water occurred between ca 9750 and 5920 yr B.P., and that the marked shallowing indicated by the mollusc deposits occurred between 5920 and 5275 yr B.P. This chronology is consistent with the fact that a sample from this unit in the Julu core has been radiocarbon-dated to 7050±110 yr B.P. Units thought to have been deposited at the same time as the final shallowing phase in two sites 200m and 20 km west of the core site contain abundant *Nymphaea* seeds and *Trapa* fruits, consistent with accumulation in very shallow water.

The overlying unit (4.0-4.2m) in the outcropping section is yellow to rusty yellow, calcareous silty clay, with discontinuous horizontal bedding. There are molluscs present. This unit is not found in the Nanwangzhuang Borehole, but can be seen in the Zhaohuotuo profile where it is ca 30 cm thick. The discontinuous nature of this unit indicates that it is unlikely to be lacustrine in origin.

The overlying unit in both the outcropping section (3.67-4.0m) and the Nanwangzhuang Borehole (4.67-5.2m) is a grey to black muddy clay, with abundant organics and containing plant roots. Mollusc shells are relatively abundant within this unit. The lithology is consistent with shallow lacustrine conditions. Samples from the top and bottom of this unit (3.67 and 4.0m respectively) have been radiocarbon dated to 1925±131 yr B.P. and 2642±96 yr B.P. This phase of shallow lacustrine conditions occurred ca 2640-1930 yr B.P.

The overlying unit (3.67-3.1 m in the outcropping section, 4.1-4.67m in the Borehole) is brown-yellow clay with horizontal bedding and thin laminations, suggesting the lake became deeper. The aquatic assemblage is characterised by Potamogetonaceae, is consistent with deepening. The ostracode assemblage is not described. By interpolation between the radiocarbon date from the top of the underlying unit and assuming that the top of the outcropping section is ca 200 years old, this phase occurred ca 1930-1660 yr B.P.

The overlying unit (3.1-0.85 m in the outcropping section, 1.85-4.1m in the Borehole) is brown silty clay with horizontal bedding and lamination. The increase in silt content suggests the lake became slightly shallower. However, the ostracode assemblage is dominated by *Candona*, so the lake must have been relatively deep. The absence of aquatic pollen is consistent with this interpretation. This phase occurred ca 1660-1070 yr B.P.

The uppermost unit (0-0.85 m in the outcropping section, 1-1.85 m in the Borehole) is brown massive clay and silty clay, with abundant plant roots and mollusc shells. The absence of lamination, and the abundance of plant roots and molluscs, suggests the lake became shallower. The ostracode assemblage is dominated by *Ilyocypris*, consistent with decreased water depth. This phase occurred 1070-200 yr B.P.

In the status coding, very low (1) is indicated by dry conditions in the central part of the lake basin; low (2) by calcareous nodules and very abundant molluscs, with *Ilyocypris*; moderately low (3) by organic-rich muddy clay, with abundant molluscs, or organic silty clay with *Ilyocypris*; intermediate (4) by silty clay, with *Candona* or with deepwater aquatic pollen (Potamogetonaceae); high (5) by clay, displaying horizontal bedding and lamination, with an aquatic assemblage characterised by Potamogetonaceae and the absence of Alismataceae, and with *Candona*.

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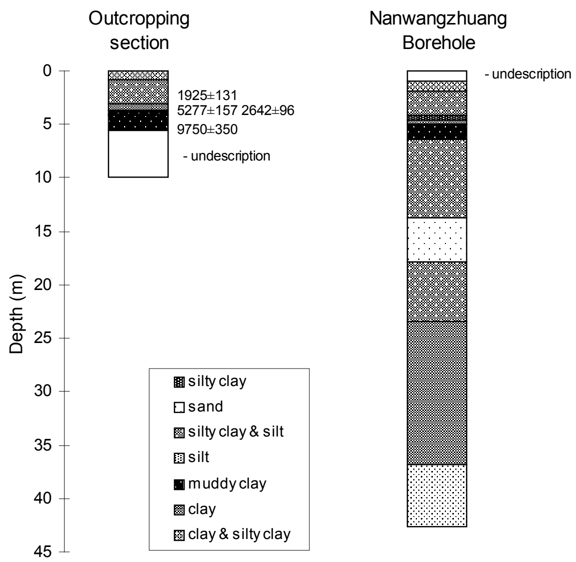

Radiocarbon dates

9750±350	5.6 m, organic components from outcropping section
7050±110	organic material, Julu Profile
5277±157	4.2 m, organic components from outcropping section
2642±96	4.0 m, organic components from outcropping section
1925±131	3.67 m, organic components from outcropping section

Note: the date from the Julu Profile is not used in assessing dating control.

Coding

ca 18,000 yr B.P.	very low (1)
pre 9750 yr B.P.	intermediate (4)
9750-5920 yr B.P.	high (5)
5920-5275 yr B.P.	low (2)
5275-2640 yr B.P.	very low (1)
2640-1930 yr B.P.	moderately low (3)
1930-1660 yr B.P.	very high (5)
1660-1070 yr B.P.	intermediate (4)
1070-200 yr B.P.	moderately low (3)
200-0 yr B.P.	very low (1)

Preliminary coding: 01-02-1999
Final coding: 14-03-1999
Coded by BX, GY and SPH

3.4. Xingkai Lake (Khanka Lake), Heilongjiang Province

Xingkai Lake (45° 10' N, 132° 10' E, 69 m a.s.l.) is a freshwater lake which lies across the boundary between China and Russia. The lake has a mean depth of 6.2m, a maximum depth of 10m (Wang et al., 1987) and an area of 4800 km^2. The lake is fed by three rivers flowing into the lake from the west and southwest, and overflows via the Songacha River. The lake basin is of tectonic origin, but tectonic movements have been weak during the late Pleistocene and do not appear to have influenced the lake-level history (Qiu et al., 1988).

Qiu et al. (1988) have reconstructed the history of lake-level changes within this basin, on the basis of geomorphic studies of the relict beach ridges found on the gently-sloping lacustrine plain on the northern (i.e. Chinese) side of the lake. There are five beach ridges. Unfortunately, the height of the top of the ridges is not given in the original publication, and there is no indication of the change in elevation across the relict lacustrine plain. Thus, it is not possible to determine the absolute lake level at the time of formation at each of the ridges. Changes in relative water depth are established on the assumption that distance from the modern shoreline is a reasonable measure of changes in lake depth. Two of the ridges have been dated, using TL, and a further TL date has been obtained from associated lacustrine sediments. A single radiocarbon date, on archaeological material associated with the youngest ridge provides further chronological control.

The fifth (outermost) ridge, lies ca 16 km to the northeast of the modern lake shoreline. The ridge is ca 85 km long, 70-150 m wide and 6-10m high. The deposits consist of well-sorted, well-rounded fine and medium sand, and show clear, inclined bedding. Thus, both the lithology and the structures are consistent with shoreline (beach) deposits. Accumulation of beach sand was apparently intermittent, since at least 12 distinct palaeosols are recognised within the deposits. A sample from a depth of 1.15m, taken from the beach deposits between the youngest and the next youngest palaeosol, was TL-dated to 63,900±3,190 yr B.P. The fifth beach ridge thus indicates that the lake was considerably more extensive than today before ca 63,000 yr B.P. The thickness of the beach deposits suggests that the lake was more extensive for a considerable period of time, although the development of palaeosols within the deposits clearly show that it underwent minor fluctuations in depth during this period.

The fourth beach ridge is of limited extent (9.5 km) and lies ca 4km closer to the lake than the firth ridge. The ridge is only 30-50m wide and 1.5-3.0m high. The deposits consist of fine to medium sand, but the stratigraphy has not been investigated in detail. The ridge has not been dated. Qiu et al. (1988) suggest it marks an interval of lower lake level subsequent to the formation of the fifth ridge, and that the poor development of the feature indicates that this lake stand was relatively short-lived. They suggest that the fourth ridge could have been formed between 50,000 and 35,000 yr B.P. An alternative explanation for its limited extent, small height and poor preservation is that the feature was formed before the deposition of the outermost beach ridge and was subsequently partially destroyed and modified as lake level rose to the elevation marked by the fifth beach ridge.

A 2m-deep profile from the lake plain deposits ca 5 km to lakeward of the fifth beach ridge consists of grey to brown sands. A sample from 1.75m was TL-dated to 63,000±3,100 yr B.P. The exact stratigraphic relationship between these lacustrine deposits and the various beach ridges is unknown, but the dating suggests that the lake was still considerably more extensive than today at 63,000 yr B.P.

The third beach ridge is a well-developed feature, lying between 5-10 km away from the modern shoreline. The ridge is 76 km long, 60-100m wide and 2-3m high. The deposits consist of well-sorted, well-rounded fine and medium sands. The ridge is undated, although it clearly postdates formation of the outermost ridges. Qiu et al. (1988) speculate that it was formed during the last interstadial.

The second beach ridge only occurs in the northeast quadrant of the basin. At its eastermost end it is about 2-3 km away from the modern lakeshore, but it gradually gets closer to and finally merges into the first beach ridge after a distance of ca 27 km. The ridge is ca 10-20m wide and 12-14m high at its maximum extent. The deposits consist of well-sorted, well-rounded fine and medium sand. There are no dates on the feature.

The first beach ridge appears to mirror the position of the modern shoreline. The ridge is 87.5 km long, 10-20m wide and 6-10m high. The deposits consist of well-sorted, well-rounded beach sands. There are six sand units, separated by palaeosols. A sample from one of the sand units (2.01m) was TL-dated to 12,190±610 yr B.P. A sample from archaeological material overlying the beach ridge was radiocarbon-dated to 5430 yr B.P. The first beach ridge can be interpreted as indicating a period of higher-than-present lake levels during the deglacial and early- to mid-Holocene. At ca 12,200 yr B.P., the lake level must have been ca 4m higher-than-today, and lake levels must subsequently have risen in order to form the overlying 2m of beach deposits. Cessation of beach ridge formation had occurred at least by 5430 yr B.P., though it could have taken place earlier.

There is no geomorphic evidence for lake levels higher than today after ca 5400 yr B.P.

In the status coding, low (1) is indicated by modern lake level and the absence of sand ridges above modern lake level in the late Holocene; intermediate (2) by the lake beach deposits of the first beach ridge, which indicate lake levels up to 10m higher than present but no great increase in lake area; high (3) by beach ridges indicating the lake was spatially more extensive than today.

References

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Dating

63,900±3190	TL-date, sand, 1.15m below top of fifth (outermost) sand ridge
63,000±3100	TL-date, sand, 1.75m below surface, profile, lake plain ca 5km SW of ridge 5
12,190±610	TL-date, sand, 2.01m below top of first (innermost) sand ridge
5430±	^{14}C date, archaeological material on top of first sand ridge

Coding

pre-64,000 – 63,000 yr B.P.	high (3)
pre-12,200 yr B.P.	intermediate (2)
12,200-5400 yr B.P.	intermediate (2)
5400-0 yr B.P.	low (1)

Preliminary coding: December 1998
Second coding: January 1999
Final coding: March 1999
Coded by BX and SPH

3.5. Longquanhu, Hubei Province

Longquanhu (30°52'N, 112°2'E, ca 150 m a.s.l.) was a swamp before the 1950's, and was artificially drained for farmland afterwards (Li et al., 1992). The former swamp had an area of approximately 500 m^2. There is no surface inflow and the catchment area is confined to the small hills around, which rise to ca 200 m a.s.l. The annual mean precipitation is ca 900-1400 mm.

Two cores were taken close together from the central part of the palaeolake (Li et al., 1992; Liu, 1993). One core (LC1), 4.0 m long, provides a lithological record back to before ca 7000 yr B.P. Another core (LC2), 5.72 m long, provides a lithological and aquatic pollen record back to ca 10000 yr B.P. Changes in relative water depth are reconstructed on the basis of the changes in lithology, aquatic pollen and sedimentation rates from the two cores. The description of the lithology of core LC2 is slightly different in the text, in the table describing the radiocarbon dates and in the pollen diagrams (Liu, 1993); we have therefore erected a composite lithology that reflects the most likely interpretation of these separate descriptions, and agrees best with the lithological description of core LC1. There is only one date on core LC1, but there are 7 radiocarbon dates from core LC2 (Liu, 1993). The lowermost sample, from the basal sand layer in core LC2 (5.6-5.72m), is dated to 21910±200 yr B.P. The sedimentation rate between this date and a date on the overlying unit (5.1-5.2m), is extremely low (0.009 mm/yr), suggesting that there might be a major hiatus in deposition. Although the original author uses this basal date (Liu, 1993), we prefer to assume that it is unreliable. Thus, our chronology is based on 6 of the radiocarbon dates from core LC2 and the single date from core LC1. There are some discrepancies between the radiocarbon ages given for some samples in Li et al. (1992) and Liu (1993). These are not large enough to affect the overall chronology. We have used the dates given in the latest publication (i.e. Liu, 1993).

The basal unit of core LC2 (5.72-5.4 m) is black sand, suggesting shallow water conditions. The aquatic pollen assemblage is characterised by relatively high percentages of Cyperaceae (8-20%) and moderate values of *Typha* (<20%), consistent with moderately shallow conditions. The date of 21910±200 yr B.P. on a sample from this unit is likely to be too old. Extrapolation of the sedimentation rate (0.91 mm/yr) between the two radiocarbon dates on the overlying unit indicates this shallow-water phase occurred before 9600 yr B.P.

The overlying unit in core LC2 (5.4-5.0 m) is black silt. The finer lithology suggests the lake became deeper. No samples from this unit were analysed for pollen. Two samples from 5.1-5.2 m and 4.9-5.05 m are radiocarbon dated to 9320±215 and 9155±195 yr B.P. respectively, indicating this phase of increased water depth occurred ca 9600-9160 yr B.P.

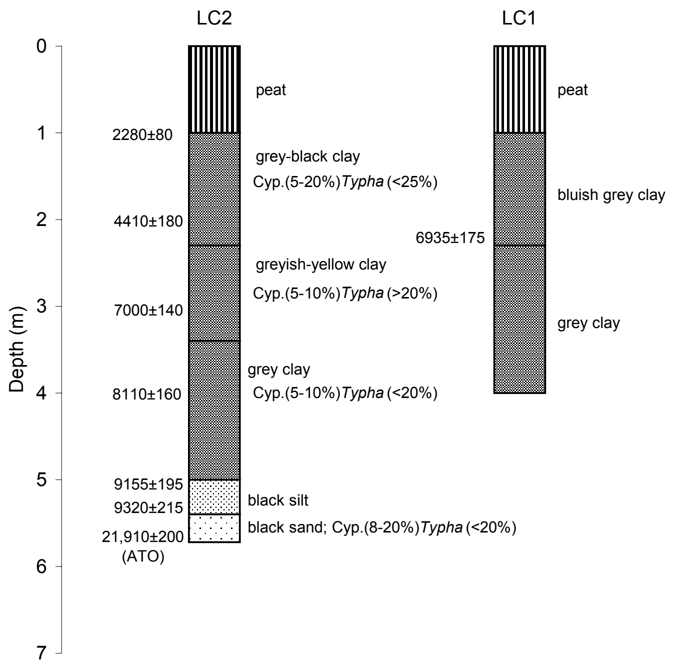
The overlying unit in core LC2 (5.0-3.4 m) is grey clay, suggesting a further deepening. The aquatic pollen assemblage is characterised by low abundance of Cyperaceae (ca 5-10%) and moderate values of *Typha* (<20%). The decrease in the abundance of Cyperaceae, compared to the basal layer, is consistent with increased water depth. A sample from 3.9-4.05 m is radiocarbon dated to 8110±160 yr B.P. Linear interpolation between this date and a date from the overlying unit, indicates that this phase of deeper water conditions occurred ca 9160-7440 yr B.P.

The overlying unit in core LC2 (3.4-2.3 m) is greyish-yellow clay. The change in colour probably reflects increased inputs from the catchment, and suggests that the lake became shallower. Although there is no change in the abundance of Cyperaceae, there is a distinct increase in *Typha* (>20%). This change in the aquatic pollen assemblage is consistent with shallowing. A sample from 2.9-3.05 m is radiocarbon dated to 7000±140 yr B.P., indicating that this phase occurred ca 7440-5200 yr B.P. The record from core LC1 (4.0-2.3 m) during this interval also shows deposition of grey clay.

The overlying unit in core LC2 (2.3-1.0 m) is grey-black clay, while in LC1 the material is described as bluish-grey clay (2.3-1.0m). The change in colour in both cores is consistent with an increase in water depth. The aquatic pollen record from LC2 shows an increase in Cyperaceae (5-20%) and a decrease in *Typha* (<25%), consistent with increased water depth. Two samples from 1.9-2.05 m and 0.9-1.05 m in core LC2 are radiocarbon dated to 4410±180 yr B.P. and 2280±180 yr B.P. respectively, suggesting that this phase of increased water depth occurred ca 5200-2300 yr B.P.

The overlying unit in both cores (above 1m) is peat, reflecting shallowing since ca 2300 yr B.P. There is a modern soil developed in the uppermost part of this unit.

In the status coding, very low (1) is indicated by peat deposition; low (2) by black sand, with *Typha* values of <20%; intermediate (3) by black silt; high (4) by grey clay in LC1 and greyish-yellow clay in LC2, with *Typha* (>20%), very high (5) by bluish-grey clay in LC1 and grey-black clay in LC2, and *Typha* values (<25%).



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Radiocarbon dates

CG1979	21,910±200	5.6-5.72 m, silt, LC2, ATO
CG1994	9320±215	5.1-5.2 m, silt, LC2
CG1978	9155±195	4.9-5.05 m, silt, LC2
CG1993	8110±160	3.9-4.05 m, grey clay, LC2
CG1977	7000±140	2.9-3.05 m, grey clay, LC2
BK86075	6935±175	2.1-2.3 m, grey-black clay, LC1
CG1992	4410±180	1.9-2.05 m, grey clay, LC2
CG1976	2280±80	0.9-1.05 m, peat, LC2

Coding

-9600 yr B.P.	low (2)
9600-9160 yr B.P.	intermediate (3)
9160-7440 yr B.P.	very high (5)
7440-5200 yr B.P.	high (4)
5200-2300 yr B.P.	very high (5)
2300-0 yr B.P.	very low (1)

Preliminary coding: December 1998
Second coding: January 1999
Final coding: February1999
Coded by BX and SPH

3.6. Baisuhai, Inner Mongolia Autonomous Region

Baisuhai (42°35'N, 115°56'E, ca 2000 m a.s.l.), which is situated in the eastern part of Daqing Mountain, lies in a very small closed basin. The lake is fed by direct precipitation and some streams draining the catchment. The basin bedrock is Tertiary basalt.

Two cores were taken rather close together from the lake shore. The cores do not represent a transect across the lake. One core (A) is 285 cm long and the second (B) is 165 cm, and provide a lithological record back to ca 13250 and 7500 yr B.P. respectively (Cui and Kong, 1992; Cui et al., 1993). Core A also provides a record of changes in mollusc assemblages and plant macrofossils. Changes in water depth are reconstructed from changes in lithology from both cores, and changes in plant macrofossils and mollusc assemblages from core A. The chronology is based on 7 radiocarbon dates from core A (Cui et al., 1993) and 5 radiocarbon dates from core B (Cui and Kong, 1992).

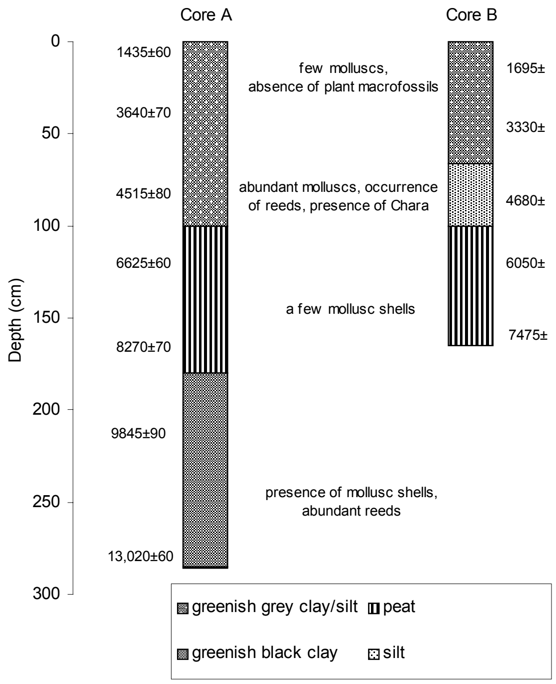
The basal unit of core A (285-180 cm) is greyish black lacustrine clay, containing shells of *Sphaerium* sp. and *Hippeutis minutus*. The plant macrofossils are dominated by reed roots and stems (including *Phragmites*), with a few *Chara*. The presence of mollusc shells and the abundance of reed macrofossils are consistent with relatively shallow water conditions. Two samples from ca 280 cm and 210 cm are radiocarbon dated to 13020±60 and 9845±90 yr B.P. respectively, indicating that this phase occurred from ca 13250 yr B.P. to 8500 yr B.P. The date of the upper boundary using the calculated sedimentation rate of 0.27 mm/yr in the overlying peat in core A is 8820 yr B.P.

The overlying unit (180-100 cm) of core A and the basal unit (165-100 cm) in core B are peat, indicating shallowing of the lake. There are only a few mollusc shell fragments present, consistent with the conversion from a lacustrine to a peat environment. Cui and Kong (1992) recognized two subunits within the peat: the lowermost characterised by the presence and the uppermost by the absence of *Sphagnum* in an otherwise relatively diverse assemblage. Cui and Kong (1992) suggest that the disappearance of *Sphagnum* indicates further shallowing of the lake consequent on swamp succession. Thus, the disappearance of *Sphagnum* is not indicative of a lake-level change and is not coded as such. Two samples from 165 and 120 cm in core A are radiocarbon dated to 8270±70 and 6625±60 yr B.P. respectively. Two samples from the depth of ca 162 and 117.5 cm in core B are radiocarbon dated to 7475 and 6050 yr B.P. respectively. The dating of the upper boundary of the peat in core A varies from 5890 (using the calculated sedimentation rate of 0.27 mm/yr between the two dates from the peat) to 4950 yr B.P. (using the sedimentation rate of 0.46 mm/yr in the overlying unit). The dating of the upper boundary in core B is between 5490 (based on the calculated sedimentation rate of 0.31 mm/yr within the peat) and 5170 yr B.P. (using the sedimentation rate of 0.30 mm/yr from the overlying two units).

The overlying unit (above 100cm in both cores) is lacustrine greenish-grey silt and clay. The change in lithology suggests that the lake became deeper. The basal part of the unit is more silty; in Core B a distinct silt-rich layer is distinguished (100-66cm). The gradual transition from silty clays to purer clays up-profile is consistent with a gradual increase in water depth. The biotic record from Core A initially shows abundant *Sphaerium* sp. and *Gyraulis alba* shells, the occurrence of reed macrofossils (including *Phragmites*) and the presence of *Chara*, consistent with relatively shallow lacustrine conditions. There is a gradual decrease in the abundance of all of these indicators, such that the uppermost part of the unit (0-30cm) is devoid of plant macrofossils and has only low abundances of molluscs. This shift is also consistent with a gradual increase in water depth. Samples from ca 80 cm, 40 cm and ca 5 cm in Core A are radiocarbon dated to 4515±80, 3640±40 and 1435±60 yr B.P. respectively. Samples from ca 85 cm, ca 44 cm and ca 18 cm in core B are radiocarbon dated to 4680 and 3330 and 1695 yr B.P. respectively. Interpolation of the sedimentation rate between the radiocarbon dates on Core A suggests that the final, deepest water phase (recorded by the absence of plant macrofossils) began ca 3010 yr B.P.

Cessation of deposition at the coring sites indicates that the lake became smaller during the last few hundred years. There are modern plant roots in the uppermost part of core A. Extrapolation of the sedimentation rate between the radiocarbon dates on the uppermost unit in Core A suggests that the lake retreated from the site ca 1120 yr B.P. The timing of lake retreat is estimated at 560 yr B.P., using the sedimentation rate on the uppermost sediment unit in Core B.

In the status coding, very low (1) is indicated by lack of deposition at the coring site; low (2) by peat; intermediate (3) by greenish-grey clay or silty-clay, the presence of mollusc shells and abundant plant macrofossils; high (4) by greenish- grey clay, with relatively few mollusc shells and absence of plant macrofossils.



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Cui HT, Wu WL, Song CQ, Wu HL (1993) Reconstruction the the Holocene environment in the Daqingshan region of inner Mongolia, In: Zhang LS (ed.) Study on the History of the Living Environment in China. Ocean Press, Beijing, China, pp. 285-295

Radiocarbon dates

13,020±60	ca 280 cm, clay, core A
9845±90	ca 210 cm, clay, core A
8270±70	ca 165 cm, peat, core A
7475±	ca 162 cm, peat, core B
6625±60	ca 120 cm, peat, core A
6050±	ca 117.5 cm, peat, core B
4680±	ca 85 cm, organic matter, core B
4515±80	ca 80 cm, organic matter, core A
3640±70	ca 40 cm, organic mud, core A
3330±	ca 44 cm, mud, core B
1695±	ca 18 cm, mud, core B
1435±60	ca 5 cm, organic mud, core A

Coding

13,250-8500 yr B.P.	intermediate (3)
8820-4950 yr B.P.	low (2)
5890-3010 yr B.P.	intermediate (3)
3010-560 yr B.P.	high (4)
1120-0 yr B.P.	very low (1)

Preliminary coding: March 1996
Final coding: February 1999
Coded by BX and SPH

3.7. Chagannur, Inner Mongolia Autonomous Region

Chagannur (43°16'N, 112°54'E, ca 920 m a.s.l.) is a salt lake in Inner Mongolia. The lake, which has an area of 21 km² today, consists of a smaller northeastern sub-basin (East Chagannur) and a larger southwestern sub-basin (West Chagannur) joined by a narrow channel (Zheng et al., 1992). Two intermittent rivers discharge into the lake, one from the northeast and the other from the southeast. The lake is also fed by direct precipitation. Chagannur is the largest of a series of 16 salt lakes (e.g. Hushunur, Wulannur, Huguonur, Halefushuyingnur, Muyingnur, Wulannur, Wenduobunur, South Chagganur) that are relicts of a formerly more extensive lake (Palaeolake Chagannur). At its maximum extent, which is thought to have been during the early to mid-Holocene, this palaeolake occupied an area of ca 2640 km² (Zheng et al., 1992). The total area of lakes in the basin today is ca 156 km². The area of the catchment is 2800 km². The basin, which is controlled by two faults running SW-NE, is of tectonic origin. The bedrock is Cretaceous sandstone and mudstone.

A 23.41 m core (Core 83-CK1), taken from near the center of the lake, provides a lithological record back to ca 20,000 yr B.P. (Zheng et al., 1992). Changes in water depth are reconstructed from changes in lithology, specifically in the relative importance of detrital and chemical deposits. The chronology is based on three radiocarbon dates from the core (Xu, 1993).

The basal unit (23.41-20.93 m) is brown silty clay with horizontal bedding. The lithology, and the preservation of horizontal bedding, suggests the lake was fresh and moderately deep (Zheng et al., 1992). By extrapolation of the sedimentation rate (2.48 mm/yr) estimated between the radiocarbon dates of 16,309±121 yr B.P. at 19.6 m and 12,554±80 yr B.P. at 10.3 m, this moderately deep-water phase occurred ca 17,850-16,850 yr B.P.

The overlying unit (20.93-20.07 m) is black mud. The increase in the organic content of the sediments and the disappearance of horizontal bedding suggests the lake became shallower, although remaining fresh. By extrapolation of the sedimentation rate from the overlying units, this phase occurred ca 16,850-16,500 yr B.P.

The overlying unit (20.07-19.71 m) is greyish-white trona. This change to chemical deposition indicates further shallowing after ca 16,500 yr B.P.

The overlying unit (19.71-19.01 m) is black mud, containing nitrate crystals. The change to predominantly detrital deposition indicates the lake became deeper. The presence of nitrate crystals, however, indicates that the increase in water depth was moderate. A sample from 19.6 m is radiocarbon dated to 16309±121 yr B.P., suggesting this phase occurred ca 16,350-16,070 yr B.P.

The overlying unit (19.01-15.71 m) is greyish-white to dark grey trona. The return to chemical deposition indicates the lake became shallower between ca 16,070 and 14,740 yr B.P. The upper part of the unit contains two thin layers of muddy clay, which could suggest short-lived intervals of increased water depth or could mark the gradual transition to the overlying, more organic unit. Unfortunately, the depth of these muddy clay layers is not given, so it is not possible to code this transition.

The overlying unit (15.71-14.61 m) is black mud. The return to detrital deposition indicates that the lake became deeper. This phase occurred ca 14,740-14,290 yr B.P.

The overlying unit (14.61-13.65 m) is greyish-white trona, with thin mud interbeds. The change to predominantly chemical deposition indicates the lake became shallower, although the continued presence of mud interbeds indicates that the change in water depth was not very large. This phase occurred ca 14,290-13,910 yr B.P.

The overlying unit (13.65-12.39 m) is black mud containing trona. The change in lithology suggests the lake became slightly deeper. This phase occurred ca 13,910-13,400 yr B.P.

The overlying unit (12.39-9.39 m) is greyish-white trona. The lower part of the unit contains three thin mud layers, suggesting a gradual transition from moderately deep to shallow water conditions. Unfortunately, the exact depths of the mud layers are not given. A sample from 10.3 m is radiocarbon dated to 12,554±80 yr B.P. By interpolation between this date and the date in the overlying unit, the upper boundary of the unit is dated to ca 10,560 yr B.P.

The overlying unit (9.39-8.04 m) is black mud with trona, indicating the lake became deeper. A sample from 8.94 m is radiocarbon dated to 9569±80 yr B.P. Assuming that the top of the core is modern, and interpolating the sedimentation rate (0.93 mm/yr) between this date and the coretop, the upper boundary of this black mud unit is dated to 8610 yr B.P.

The overlying unit (8.04-7.64 m) is greyish-white trona, indicating the lake became shallower. This phase occurred ca 8610-8180 yr B.P.

The overlying unit (7.64-7.0 m) is black mud containing nitrate crystals, indicating the lake became deeper. This phase occurred ca 8180-7490 yr B.P.

The overlying unit (7.0-6.64 m) is greyish-white trona, indicating a return to shallower conditions. This phase occurred ca 7490-7110 yr B.P.

The overlying unit (6.64-5.04 m) is black mud interbedded with thin-layers of trona. The dominance of detritic deposition indicates that the lake became deeper, although the continued presence of trona suggests that the increase in water depth was not great. This phase occurred ca 7110-5390 yr B.P.

The overlying unit (5.04-4.34 m) is greyish-white trona, with several thin mud interbeds, indicating a return to somewhat shallower conditions. This phase occurred ca 5390-4650 yr B.P.

The overlying unit (4.34-4.0 m) is black mud containing thenardite, indicating the lake became slightly deeper. This phase occurred ca 4650-4280 yr B.P.

The overlying unit (4.0-3.64 m) is light grey trona, suggesting the lake became shallower. This phase occurred ca 4280-3900 yr B.P.

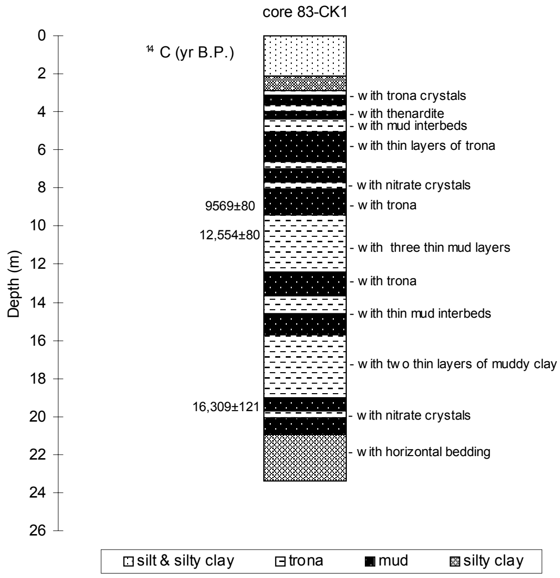
The overlying unit (3.64-3.14 m) is black mud containing trona crystals, indicating the lake became slightly deeper. This phase occurred ca 3900-3360 yr B.P.

The overlying unit (3.14-2.84 m) is grey to greyish-white trona, suggesting the lake became shallower. This phase occurred ca 3360-3040 yr B.P.

The overlying unit (2.84-2.14 m) is greyish-green clay, suggesting the lake became significantly deeper. This phase occurred ca 3040-2290 yr B.P.

The uppermost unit (2.14-0 m) is yellow brown silt and silty clay. This change in lithology suggests the lake became shallower. The origin of this unit is not clear. The original authors indicated that the unit could be fluvio-lacustrine in origin, but this implies relatively wet conditions in the recent past which is inconsistent with the observation that the lake area was substantially reduced during the late Holocene.

In the status coding, very low (1) is indicated by trona; low (2) by trona with mud layers; moderately low (3) by organic mud with trona; intermediate (4) by organic mud with nitrate crystals; moderately high (5) by organic mud with thernadite; high (6) by organic mud; very high (7) by silty clay; and extremely high (8) by clay deposition.



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Radiocarbon dates

16,309±121	ca 19.6 m, organic components
12,554±80	ca 10.3 m, organic components
9569±80	ca 8.94 m, organic components

Coding

17,850-16,850 yr B.P.	very high (7)
16,850-16,500 yr B.P.	high (6)
16,500-16,350 yr B.P.	very low (1)
16,350-16,070 yr B.P.	intermediate (4)
16,070-14,740 yr B.P.	very low (1)
14,740-14,290 yr B.P.	high (6)
14,290-13,910 yr B.P.	low (2)
13,910-13,400 yr B.P.	moderately low (3)
13,400-10,560 yr B.P.	low (2)
10,560-8610 yr B.P.	moderately low (3)
8610-8180 yr B.P.	very low (1)

8180-7490 yr B.P.	intermediate (4)
7490-7110 yr B.P.	very low (1)
7110-5390 yr B.P.	moderately low (3)
5390-4650 yr B.P.	low (2)
4650-4280 yr B.P.	moderately high (5)
4280-3900 yr B.P.	very low (1)
3900-3360 yr B.P.	moderately low (3)
3360-3040 yr B.P.	very low (1)
3040-2290 yr B.P.	extremely high (8)
2290-0 yr B.P.	very high (7)

Preliminary coding: February 1999
Final coding: March 1999
Coded by BX, GY and SPH

3.8. Erjichuoer, Inner Mongolia Autonomous Region

Erjichuoer (45°14'N, 116°30'E, 829.2 m a.s.l.) is a closed sulfate-type salt lake in Inner Mongolia. The lake has an area of 10 km² and is surrounded by playa deposits (chiefly mirabilite) extending over a further 16 km². The water depth varies from 0.05 m to 0.3 m seasonally. The lake is fed by direct precipitation and local ground water. The annual precipitation is ca 250-300 mm, and the mean annual evaporation is high (up to 2000 mm/yr). The basin has an area of 700 km² and is of tectonic origin. There is no evidence of recent tectonic activity. The bedrock includes Tertiary mudstones, Lower Cretaceous sandstone and mudstones, and Permian metamorphics. Quaternary sediments are represented by between 80-110 m of sandstone and sandy conglomerate (Zheng et al., 1992).

A series of three inset lacustrine terraces around the margin of the modern lake indicates phases of higher lake levels in the past. The oldest terrace, which overlies Tertiary mudstone bedrock, consists of sand and gravel deposits. The second terrace consists of fine beach sand. The youngest terrace is formed from sandy clay deposits. Facies equivalents of the two younger terraces can be found underlying the salt-dominated units that now form the lake bed. Unfortunately, there are no dates on the terraces.

The chronology of past lake-level changes in the Erjichuoer basin can be reconstructed from a 11.04 m-long core (Erjichuoer 83-CK_1), taken from the central part of the modern lake, which provides a lithological record back to ca 15,160 yr B.P. (Zheng et al., 1992). Changes in water depth are based on changes in lithology, and specifically changes in the importance of detritic and chemical deposition. There are two radiocarbon dates from the core.

The basal unit (11.04-10.0 m) is grey muddy silt and fine sand, and is likely the facies equivalent of the middle terrace deposit. The lithology is consistent with a moderately shallow, freshwater lake (Zheng et al., 1992). By extrapolation of the sedimentation rate (1.05 mm/yr) between the two radiocarbon dates of 12,074±139 and 7,503±864 yr B.P. at 7.8 m and 3.0 m respectively, this phase occurred ca 15,160-14,170 yr B.P.

The overlying unit (10.0-7.65 m) is grey to greyish-green clay, and is likely the facies equivalent of the youngest terrace deposit. The change in lithology suggests the lake became deeper. By extrapolation of the sedimentation rate between the available radiocarbon dates, this phase occurred ca 14,170-11,930 yr B.P.

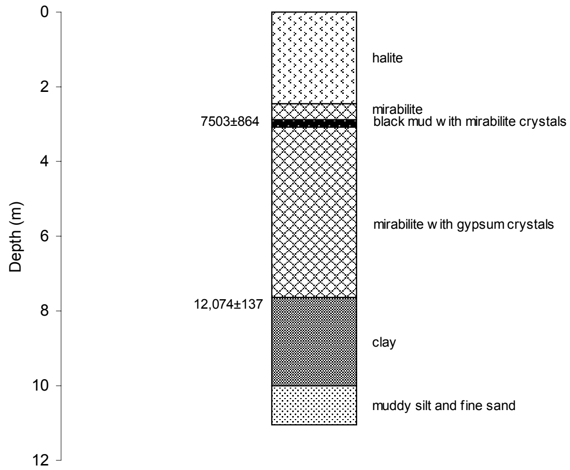
The overlying unit (7.65-3.09 m) is grey mirabilite, containing gypsum crystals. The change to predominantly chemical deposition suggest the lake became considerably shallower. This phase occurred ca 11,930-7590 yr B.P. The timing of this interval is well constrained, since the radiocarbon dating samples were taken from close to the lower and upper boundaries of the unit.

The overlying unit (3.09-2.89 m) is black mud, containing some mirabilite crystals. This change in lithology suggests an increase in water depth. A sample from ca 3.0 m is radiocarbon dated to 7503±864 yr B.P. By interpolation of the sedimentation rate (0.40 mm/yr) between this date and the top of the core, which is assumed to be modern, this phase of increased water depth occurred ca 7590-7230 yr B.P.

The overlying unit (2.89-2.47 m) is mirabilite, indicating a return to shallower conditions. This phase occurred ca 7230-6180 yr B.P. The lake area at this stage can be estimated from the extent of mirabilite across the basin to have been ca 26 km².

The uppermost unit (2.47-0 m) is halite with thin mud interbeds, suggesting further shallowing after ca 6180 yr B.P. Halite is being deposited in the centre of the basin today.

In the status coding, very low (1) is indicated by halite; low (2) by mirabilite; intermediate (3) by black mud containing mirabilite or gypsum crystals; high (4) by silt and fine sand; very high (5) by grey to greyish-green clay.



Reference

Zheng XY, Zhang MG, Dong JH, Gao ZH, Xu C, Han ZM, Zhang BZ, Sun DP, Wang KJ (1992) Salt Lakes in Inner Mongolia. Science Press, Beijing, 1-296 (in Chinese)

Radiocarbon dates

12,074±139	ca 7.8 m, clay
7503±864	ca 3.0 m, mud

Coding

15,160-14,170 yr B.P.	high (4)
14,170-11,930 yr B.P.	very high (5)
11,930-7590 yr B.P.	low (2)
7590-7230 yr B.P.	intermediate (3)
7230-6180 yr B.P.	low (2)
6180-0 yr B.P.	very low (1)

Preliminary coding: February 1999
Final coding: March 1999
Coded by BX, GY and SPH

3.9. Hulun Lake, Inner Mongolia Autonomous Region

Hulun Lake (48°31'-49°20'N, 116°58'-117°48'E, 545 m a.s.l.) is the fifth largest lake in China, and also the furthest north of the large lakes of China. The lake covers an area of 2339 km², and has a maximum depth of ca 8 m. The water salinity is 1.165 g/L. The lake is fed by direct precipitation and over 80 rivers, of which the largest three are the Crulen, Orshun-Gol and Hailaer Rivers. The Crulen River originates in Mongolia, and the Orshun-Gol River flows from the nearby Buir-nur lake on the Sino-Mongolian border. There is an outflow from Hulun Lake via the Erguna River. The regional structure, which belongs to the New Huaxia system, is controlled by 2 NNE faults. The outcropping strata are Mesozoic and upper Tertiary sandstone, mudstone, and conglomerate with interstratified coal measures. Quaternary fluvial and lacustrine deposits are widely distributed over the basement rocks within the basin and in the valley.

Unlike most of the lakes in China, Hulun lake has been characterised by a general rise of lake level during this century. The lake nearly dried out in 1900 A.D. according to historical records, and lake level rose between 1900 to 1962 from ca 536.5 to 545.0 m a.s.l. The lake level decreased to 542.9 m a.s.l. between 1962 and 1982 due mainly to the construction of a dam on the Dalanerlemu River, which connects the lake with Hailaer River. The lake level rose again during the 80's and the present level is ca 545 m a.s.l. (Wang et al., 1995).

There are no continuous lacustrine terraces around Hulun Lake. However, isolated lacustrine sediments are found in several locations at elevations of between 550 m and 560 m (i.e. between 5 and 10m above the level of modern lake). Littoral swamp deposits have been found at an elevation of 550 m (+5 m above modern lake level) in the Orshun-Gol River valley. These deposits have not been radiocarbon dated. Lacustrine shell deposits were also found at an elevation of 550 m in a section near Hangyun Station, which is located to the north of and not too far away from the modern lake. A shell from this section has been radiocarbon dated to 4790±100 yr B.P. Lacustrine silty mud with plant roots is found at an elevation of 15 m above modern lake-level in the Balongsabo section, which is located to the southwest of the modern lake. There are 5 cycles of interbedded grey lacustrine silty mud and greyish yellow nearshore sand in the Balongsabo section (Wang et al., 1995). Charcoal from the uppermost mud has been radiocarbon dated to 11,410±210 yr B.P. Thus, isolated lacustrine deposits provide evidence of at least two phases when the lake was deeper than present: ca 11500 yr B.P. (+15 m) and ca 4800 yr B.P. (+5m) (Wang et al., 1995). The evidence from Balongsabo section suggests that the earliest of these high lake-level phases was preceded by at least four other lake transgression of a similar magnitude.

A 14.6 m profile from the Donglutian open coal mine, which is situated to the north of the modern lake, with an upper elevation of ca 545 m a.s.l., provides a sedimentary record for the period ca 34,000-3000 yr B.P. (Wang et al., 1995). There are 11 radiocarbon dates in this profile, ten of which appear to be in a consistent sequence, and are consistent with the other dates from the basin. One age (ATO, 91HLD52) on the charcoal from the unit with ancient-human activities might be contaminated, and is not used to erect a chronology for the sedimentary sequence. Changes in lake level prior to 3000 yr B.P. can be reconstructed on the basis of comparison of the geomorphic evidence and the record of changes in sedimentology, diatom and some aquatic pollen assemblages from the coal mine profile. The chronology is based on 10 radiocarbon dates from the Donglutian coal mine profile, and 3 dates on other deposits within the basin (Wang et al., 1995; Li, 1984).

The basal unit (14.6-12.5 m) in the Donglutian profile is fluvial gravel, directly overlying Jurassic bedrock. *Mammuthus* and *Coelodonta* skeletons and a tree stump were found at the contact. A *Mammuthus* coprolite was radiocarbon dated to 33,760±1700 yr B.P. and the stump was dated to 28,900±1300 yr B.P. (Li, 1984). A peat bed to the southwest of the profile (Gushan section), which shows a facies transformation to the gravel layer, has been radiocarbon dated to 19,900±575 yr B.P. This suggests that the basal unit in the coal mine profile formed during the last glaciation. According to the coal mine core survey, these fluvial/alluvial sandy gravel deposits are extensively distributed in the basin, suggesting the Hulun basin was not occupied by a lake during the last glaciation.

The overlying unit (12.5-9.24 m) in the Donglutian profile is a laminated, silty clay of lacustrine origin. A sample from near the base of the unit (12.15m) was radiocarbon-dated to 12,700±230 yr B.P. The abrupt change in lithology may indicate a rapid transition to a deep lake at ca 12,850 yr B.P. or may indicate a sedimentary hiatus covering the deglacial period. Diatoms are very abundant (3.1-4.6 million per gram, with a maximum of 9.2 million per gram). The assemblage is dominated by *Melosira granulata*. The percentage of planktonic diatoms (*Cyclotella comta* and *Coscinodiscus lacustris*, 1.1 million per gram and 22 thousand per gram respectively) is relatively high throughout this unit, consistent with deep water. However, changes in the diatom assemblages appear to indicate fluctuations in water depth. Between 12.50-12.45 m planktonics dominate the assemblage. Between 12.45-10.85 m, although the diatom abundance remains high (3.9 million per gram), epiphytic species (*Fragilaria* spp) became more abundant and *Melosira* was present. This change in the diatom assemblage suggests the lake became somewhat shallower. Planktonic diatoms became dominant again in the interval between 10.85-9.24 m, indicating an increase in water depth. Samples from 12.15, 9.75 and 9.47 m have been radiocarbon dated to 12,700±230 yr B.P. (91HLD6), 11,750±550 yr B.P. (91HLD36) and 11,300±225 yr B.P. (91HLD33) respectively. On the basis of these dates, deep water conditions occurred between 12,850-12,830 yr B.P., the lake became shallower between 12,830-12,185 yr B.P. and deeper again between 12,185-11,200 yr B.P. The final phase of deep-water lacustrine deposits from the coal mine profile can be correlated with the lacustrine deposits dated to ca 11,500 yr B.P. in the Balongsabo profile. This suggests that the lake level was at least 15 m higher than the present and the lake was three times as large as the modern lake (Wang et al., 1995). The evidence for at least one phase of shallower and one of deeper water conditions before this is consistent with the evidence for fluctuations in lake level recorded in the Balongsabo profile.

The overlying unit (9.24-8.35 m) is grey silty clay, interbedded with thin layers of pale grey fine sand, and some thin layers of gyttja containing plant remains. The deposits are characterised by vein, lenticular and hummocky bedding structures, and the sands are well-sorted and rounded, suggesting the unit was deposited in a littoral environment. Diatoms are abundant with peak values of 9.4 million per gram, and the assemblage is dominated by epiphytic species (*Fragilaria* spp.), consistent with shallowing. On the basis of interpolation between the radiocarbon dates on over- and underlying units, this shallowing phase occurred between 11,200-10,900 yr B.P.

The overlying unit, 8.35-7.76 m, is pale grey fine sand, with climbing ripple (wandering and homophase), parallel and cross vein, lenticular bedding. The lithology and depositional structures suggest a beach or delta front environment, and indicate that the lake became shallower. Diatoms are absent, consistent with further shallowing between 10,900-10,600 yr B.P.

The overlying unit (7.76-6.80 m) is grey silty clay, interbedded with thin layers of fine sand, and with some thin layers of gyttja containing plant remains. This change in lithology suggests a return to nearshore deposition and indicates the lake level rose. The diatom abundance reaches 15 million per gram. Epiphytic *Fragilaria*, and *Melosira granulata* co-dominate between 7.76-7.30 m, then planktonic diatoms (*Coscinodiscus lacustris* and *Cyclotella comta*, *C.stelligera*) with the abundance of 6.6 million per gram and *Melosira granulata* co-dominate between 7.30-6.88 m, while *Fragilaria* decrease distinctively. This shift in the diatom assemblages is consistent with a gradual increase in water depth. A sample from 6.9 m is dated to 10,280±265 yr B.P., suggesting the first deep phase of this unit occurred in ca 10,600-10,450 yr B.P. and the lake became deeper ca 10,450-10,300 yr B.P.

The base of the overlying unit has a distinct scour face, overlapped by well-sorted greyish yellow coarse sand containing gravel and abundant stone artifacts, earthenware, shell and brick fragments. The abundance of human artefacts indicates that this was an occupation site, and shows that the lake level dropped after ca 10,300 yr B.P.

The overlying unit (6.80-4.20 m) is light greyish yellow fine sand, with tabular, trough cross-bedding, parallel bedding and aeolian-infill structures. The steep foreset bedding is characteristic of aeolian sand. There are no diatoms in the lower part of the unit (6.79-5.79 m) and diatoms are scarce in the upper part (0.897 thousand- 5.0 million per gram). Those diatoms present are epiphytic and benthic species (*Fragilaria* spp, *Epithemia sorex, Epithemia zebra*, and *Navicula tuscula*, *Navicula radiosa*, and *Rhopalodia gibba*), and were likely derived from erosion of exposed nearshore sediment. Aeolian sand deposition suggests the lake remained low between ca 10,300 and 7200 yr B.P.

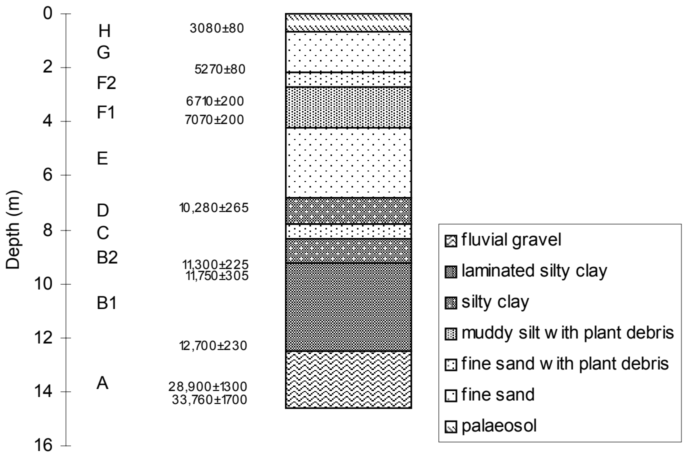
The overlying unit (4.20-2.70 m) is dark grey muddy silt, with abundant shell and plant remains. Big individual lamellibranch fossils occur concentrated at the bottom of the unit. The nature of the deposits suggests a lakeshore swamp environment. The presence of aquatic pollen such as *Typha*, *Sparganium* and *Myriophyllum* in this unit is consistent with a relatively shallow lake shore swamp environment. Diatoms are abundant (maximum 2 million per gram). The assemblage between 4.20-3.80 m is dominated by planktonic diatoms (*Coscinodiscus lacustris*) but epiphytic species (*Fragilaria*) become more common between 3.8 and 2.7m. Samples from 3.90 m and 3.60 m are radiocarbon dated to 7070±200 and 6710±200 yr B.P. respectively. This suggests that this unit was formed between ca 7200-5800 yr B.P.

The overlying unit (2.70-2.20 m) in the Donglutian profile is light greyish yellow fine sand, with thin interbeds of plant debris. The nature of the deposits suggests an aeolian dune environment, possibly with swampy interdunal areas. Thus, the lake became even shallower after 5800 yr B.P. The diatom assemblage is dominated by epiphytic species (*Epithemia sorex, E. zebra*, *Navicula tuscula*, *N. radiosa* and *Rhopalodia gibba*). This assemblage is consistent with shallowing. A sample from 2.10 m is radiocarbon dated to 5270±80 yr B.P. This phase occurred between ca 5800-4800 yr B.P.

The overlying unit (2.20-0.68 m) in the Donglutian profile is light greyish yellow fine sand, with several thin layers of light brown fine sand, rich in tabular cross-bedding, climbing ripple bedding and parallel bedding. The steep foreset bedding (near 30°) and the presence of lag gravel are characteristic of aeolian dune sand, suggesting the lake retreated from the Donglutian site ca 4800-3000 yr B.P. There are no diatoms or hygrophyte plants, consistent with this interpretation. This phase of aeolian deposition apparently occurred around the same time as shallow-water deposits, indicating a lake level ca 5m higher than the modern lake, were being formed at Hangyun Station. It is not clear how these two lines of evidence can be reconciled.

The top unit (0.68-0 m) in the Donglutian profile is a palaeosol. The presence of a palaeosol indicates the persistence of low lake levels after 3000 yr B.P.

In the status coding, very low (1) is indicated by fluvial sand or aeolian sand deposits, by palaeosol formation, or by evidence that the Donglutian site was occupied by humans; low (2) by beach sand deposits in the Donglutian profile; moderately low (3) by swamp deposits with relatively abundant epiphytic and planktonic diatoms, and the occurrence of aquatic pollen; intermediate (4) by nearshore deposits in the Donglutian profile, with abundant diatoms dominated by epiphytic species; moderately high (5) by non-laminated silty clay deposits in the Donglutian profile; high (6) by laminated silty clay with co-dominance of planktonic and epiphytic diatoms; and very high (7) by laminated silty clay with planktonic diatoms, and isolated lake deposits at an elevation of ca 15 m above modern lake level. The isolated lake deposits at +5m (Hangyun Station) cannot be correlated with the record from Donglutian and are therefore not coded. The fluctuations in lake level during the last century are not coded because they have been partly influenced by human activities.



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Radiocarbon Dates

LT2	3080±80	palaeosol, 0.34 m, Donglutian profile
91HLH2	4790±100	shell, ca 2.2 m, Hangyun Station section
LT18	5270±80	clam shell, 2.1 m, Donglutian profile
LT32	6710±200	peat, 3.6 m, Donglutian profile
LT35	7070±200	clam shell, 3.9 m, Donglutian profile
92HLD222	10,280±285	organic clay, 6.9 m, Donglutian profile
91HLD33	11,300±225	organic clay, 9.47 m, Donglutian profile
91HLB6	11,410±210	silty mud, ca 1.2 m, Balongsabo profile
91HLD36	11,750±550	organic clay, 9.75 m, Donglutian profile
91HLD6	12,700±230	organic clay, 12.15 m, Donglutian profile
91HLG	19,900±575	peat, ca 1.2 m, Gushan section
91HLD52	21,000±625	charcoal, ca. 6.7 m, Donglutian profile, ATO?
E8006	28,900±1300	paleo-tree stump, bottom of Donglutian profile
E8010	33,760±1700	coprolite of *Mammuthus*, bottom of Donglutian profile

Coding

34,000-33,500 yr B.P.	very low (1)
33,500-29,000 yr B.P.	not coded
29,000-28,000 yr B.P.	very low (1)
28,000-21,500 yr B.P.	not coded
21,500-19,500 yr B.P.	very low (1)
19,500-12,850 yr B.P.	not coded
12,850-12,830 yr B.P.	very high (7)
12,830-12,815 yr B.P.	high (6)
12,815-11,200 yr B.P.	very high (7)
11,200-10,900 yr B.P.	moderately high (5)
10,900-10,600 yr B.P.	low (2)
10,600-10,300 yr B.P.	intermediate (4)
10,300-7200 yr B.P.	very low (1)
7200-5800 yr B.P.	moderately low (3)
5800-0 yr B.P.	very low (1)

Preliminary coding: February 1996
Second coding: July 1997
Third coding: December 1998
Fourth coding: November 2000
Final coding: 23-01-2001
Coded by BX and SPH

3.10. Jilantai, Inner Mongolia Autonomous Region

Jilantai (39°45'N, 105°42'E, ca 1022.6 m a.s.l.) is a saline playa-lake in a closed basin. The playa-lake has an area of 17.81 km². The playa-lake is only fed by direct precipitation and local groundwater. Since the annual precipitation is ca 40-150 mm and the mean annual evaporation is ca 2800-4000 mm, the basin is generally dry and indeed the playa surface is currently being buried under sand. However, some highly saline water occurs in the central lowest part of the basin in years when precipitation is abundant (Geng and Cheng, 1990). The basin, which is of tectonic origin, has an area of over 2000 km². The bedrock is predominantly Palaeozoic metamorphics, with some Tertiary sandstone and mudstone in the northwestern part of the basin, and Jurassic/Cretaceous and Tertiary sandstone and mudstone in the east (Zheng et al., 1992).

The stratigraphy of the playa-lake deposits has been reconstructed from a northwest-southeast transect of 7 cores (150/CK85, 150/CK53-3, 122/CK69, 134/CK69, 118/CK85, 118/CK53-1, 114/CK49) across the basin (Geng and Cheng, 1990). Although the deposition is not perfectly regular, due to shifts in the focus of deposition and/or erosion of older sediments, these cores show a clear sedimentary sequence. The basal deposits are silty clay, overlain by clay. The detrital nature of these sediments indicates the existence of a freshwater lake. The transition from silty clay to clay indicates an increase in water depth. The overlying unit, which only occurs in the centre of the basin, is mirabilite. The shift to chemical deposition and the extent of the deposit indicate that the lake was very shallow and of limited extent. A return to wetter conditions is marked by renewed deposition of silty clay over much of the basin. A lense of fine sand associated with this silty clay deposit may mark the location of the shoreline during this phase of deeper water conditions. Shallowing is marked by a return to mirabilite deposition. A further relatively deepwater phase is indicated by silty clay deposition. A return to shallow-water conditions is indicated by the deposition of gypsum and halite. Further shallowing is shown by deposition of pure halite. However, slightly wetter conditions may be indicated by the existence of two lenses of more gypsum-rich halite within this unit. The uppermost deposits are halite. Geng and Cheng (1990) suggest these deposits were formed during the Holocene. Unfortunately, there are no radiocarbon dates on these seven cores.

Geng and Cheng (1990) indicate that this transect stratigraphy shows that the Jilantai lake has evolved during the Holocene from a freshwater lake, depositing predominantly detritic material, to a saline lake dominated by sulphate, sulphate-halite and halite salts progressively (Geng and Cheng, 1990). The extent of the lake at various times during the Holocene can therefore be reconstructed from the extent of different types of salt deposits. Thus, the extent of freshwater lacustrine deposits indicates that the lake area was > 600 km² in the early Holocene. Abundant snail shells found in coarse beach sands marking the shoreline of this freshwater lake 1 km to the north of the modern playa-lake and 2km east of the modern playa-lake have been dated to 9959±130 yr B.P. and 9940±130 yr B.P. respectively (Geng and Cheng, 1990). The extent of sulphate (mirabilite) deposits indicates that the lake area was ca 224 km² at the time of their deposition (assumed to be in the early mid-Holocene), the extent of sulphate-halite (gypsum-halite) deposits indicates the lake had an area of ca 102.4 km² probably sometime in the late mid-Holocene, and the extent of halite deposits indicates the lake had an area of 55 km² in the late-Holocene (Geng and Cheng, 1990).

The stratigraphic sequence shown in the transect cores is also shown by the record of a 17.93 m-long core (core 83-CK_1), taken from near-center of the lake (Zheng et al., 1992). There are two radiocarbon dates of 13,709 and 9782 yr B.P. at ca 14.4 m and 10.4 m respectively from this core.

The basal unit (17.93-16.71 m) in core 83-CK_1 is red to reddish-brown clay. The overlying unit (16.71-14.91 m) is grey well-sorted silt. The high debris content (>90%) of both units suggest the lake was relatively fresh. These detritic units probably correspond to the basal two units (silt clay and clay respectively) as described in the transect cores. By extrapolation of the calculated sedimentation rate (1.02 mm/yr) between the two radiocarbon dates, this phase of freshwater conditions occurred ca 17,170-14,210 yr B.P.

The overlying unit (14.91-10.30 m) in core 83-CK_1 is greyish-yellow to yellowish-brown silty clay with platy gypsum crystals. The decrease in the detritic content (55-90%) and the occurrence of gypsum crystals suggest the lake became shallower. This unit probably corresponds with the lowermost mirabilite unit in the transect cores. This shallow-water phase occurred ca 14,210-9680 yr B.P.

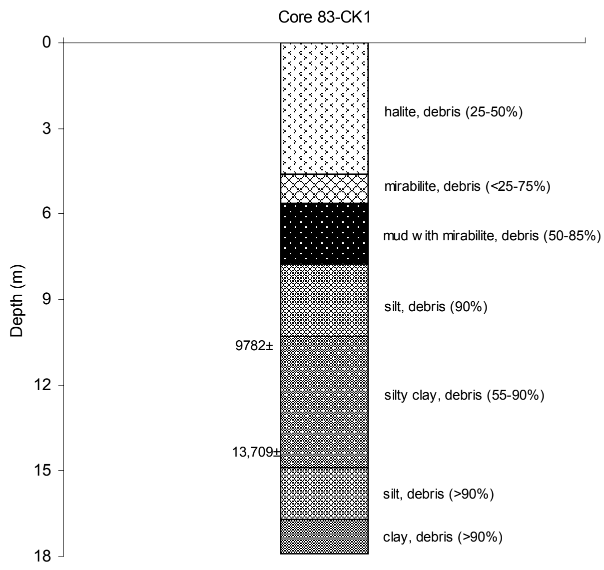
The overlying unit (10.30-7.76 m) in core 83-CK_1 is grey to greyish-green silt. The increase in the detritic content (ca 90%) suggests the lake became deeper. This unit probably corresponds to the silt clay unit which overlies the lowermost mirabilite unit in the transect cores. Using the calculated sedimentation rate of 1.02 mm/yr, this interval of increased water depth occurred ca 9680-7200 yr B.P. These dates are consistent with the early Holocene dates on the snail-containing beach deposits that were formed during the last phase of freshwater conditions in this basin.

The overlying unit (7.76-5.61 m) in core 83-CK_1 is black mud with mirabilite and gypsum. The decrease in the detritic content (50-85%), and the occurrence of mirabilite (10-60%) suggest the lake became shallower. The unit corresponds to the uppermost mirabilite unit in the transect cores. This phase occurred ca 7200-5280 yr B.P.

The overlying unit (5.61-4.61 m) is mirabilite, with some silt and gypsum. The presence of silt suggest that the lake became deeper. This unit is probably equivalent to the silt clay unit recorded in the transect cores. This phase of slightly increased water depth occurred ca 5280-4340 yr B.P.

The uppermost unit (4.61-0 m) is predominantly halite. This change in lithology indicates that the lake became shallower. The lowermost part of the unit contains some detrital material and mirabilite, while the uppermost part of the unit contains gypsum crystals and then subsequently becomes pure halite. This unit probably corresponds to the uppermost gypsum-halite and halite units in the transect cores, assumed to have been formed in the late Holocene. This phase of predominantly chemical deposition, and shallow lake conditions occurred ca 4340-0 yr B.P. The maximum water depth during this interval was 2m and the maximum area of the lake, estimated from the extent of the halite deposits, was ca 55 km^2 (Geng and Cheng, 1990).

In the status coding, very low (1) is indicated by halite deposition; low (2) by mud clay with mirabilite and gypsum; intermediate (3) silt clay with mirabilite; and high (4) by silt or silt clay, with a high detritic content (90%) and no salts.



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Radiocarbon dates

13,709±	14.4 m, organic components, core 83-CK_1
9959±130	snail shells, beach deposits east of the modern lake
9940±130	snail shells, beach deposits north of the modern lake
9782±	10.4 m, organic components, core 83-CK_1

Coding

17,170-14,210 yr B.P.	high (4)
14,210-9680 yr B.P.	intermediate (3)
9680-7200 yr B.P.	high (4)
7200-5280 yr B.P.	low (2)
5280-4340 yr B.P.	intermediate (3)
4340-0 yr B.P.	very low (1)

Preliminary coding: February1999
Final coding: 14-03-1999
Coded by BX, GY and SPH

3.11. Xidadianzi, Jiling Province

Xidadianzi (42°20'N, 126°22'E, ca 614 m a.s.l.) is a peat-bog in the Changbai Mountains of northeastern China. The peat-bog covers an area of 1.1 km², and is surrounded by small hills which have a mean altitude of ca 800 m a.s.l. The basin is of volcanic-tectonic origin and the bedrock is volcanic. The annual mean temperature of this region is 2.5 ^{o}C, and the region is characterized by a monsoonal climate (Sun and Yuan, 1990).

A northeast-southwest transect across the peatbog (Sun and Yuan, 1990) shows that the peat is underlain by lacustrine sediments. The average depth of peat across this transect is about 5-6m. The interface between the lacustrine deposits and the bedrock is very uneven. Two cores (Core X_1 and X_1C) taken close to the line of the transect have been described in detail (Sun and Yuan, 1990). These cores appear to have come from a deep basin within the bedrock, and thus to have an expanded record of lacustrine sedimentation. Core X_1 is 12 m long and X_1C is 13.5 m long. The two cores were taken very close together and display the same lithology (Sun and Yuan, 1990). A further two cores (X_2 and X_3) which are 3.2 m and 3.7 m long respectively, were taken from the northern part of the basin. There is no description of the lithology of these two cores. A single radiocarbon date was obtained from core X_3, but this date is believed to be too old by comparison with the pollen chronology. Core X_1 and the bottom 1.5 m of core X_1C was analysed for pollen. The combined record from core X_1 and X_1C provides a lithological and aquatic pollen record back to ca 10,220 yr B.P. There are 5 dates from core X_1C. The sample from 13.50-13.38 m is younger than the overlying dates and is not used to erect the chronology. Thus, the chronology is based on the other four radiocarbon dates in core X_1C.

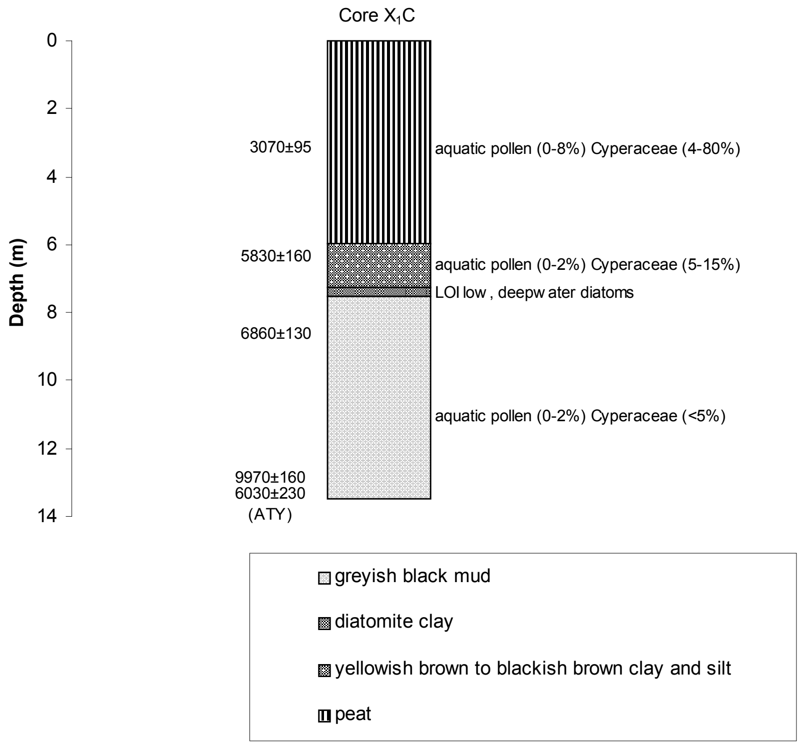
The basal unit (13.5-7.5 m in core X_1C, 12.0-7.5m in Core X_1) is greyish-black mud, containing clay and silt, and some plant debris. This unit is lacustrine in origin. The low abundance of aquatic pollen and Cyperaceae (<5%) is consistent with moderately deep water. Two samples from 13.25-13.0 m and 8.50-8.25 m are radiocarbon dated to 9970±160 and 6860±130 yr B.P. respectively. By the extrapolation of the sedimentation rate of (1.53 mm/yr) between these two dates, this phase occurred ca 10220-6290 yr B.P. Interpolation between the uppermost date and a date of 5830±160 yr B.P. from a depth of 6.25-6.0m, suggests the transition to the overlying unit occurred ca 6420 yr B.P.

The overlying unit (7.5-7.24 m) is light yellow diatomite clay, suggesting an increase in water depth. This is consistent with the fact that the loss-on-ignition is very low (Sun and Yuan, 1990). The diatom assemblage is dominated by *Fragilaria construens* var *venter* and var *binodis*, both of which are deepwater species. Extrapolation of the sedimentary rate in the underlying unit, indicates that this deepwater phase occurred ca 6290-6120 yr B.P. Interpolation between the radiocarbon dates from 8.25-8.50m and 6.0-6.25m indicates that this interval occurred between 6420 and 6300 yr B.P.

The overlying unit (7.24-5.95 m) is yellowish-brown to blackish-brown clay and silt with plant debris, and contains several thin layers of gravel and volcanic ash. The change in lithology suggests that the lake became shallower. The low abundance of aquatic pollen and Cyperaceae (5-15%) is consistent with relatively shallow conditions. A sample from 6.25-6.0 m is radiocarbon dated to 5830±160 yr B.P., indicating that the transition to the overlying unit occurred ca 5800 yr B.P.

The overlying unit (5.95-0 m) is peat, suggesting shallowing. The increase in the abundance of aquatic pollen and Cyperaceae (4-80%) is consistent with shallowing. Peat deposition has occurred continuously since ca 5800 yr B.P., suggesting that this change reflects climatic conditions rather than simply hydroseral development.

In the status coding, low (1) is indicated by peat deposition; intermediate (2) by clay and silt deposition; high (3) by diatomite clay.



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Radiocarbon dates

12,510±150	3.1-3.2 m, X_3 (ATO)
9970±160	13.25-13.0 m, silt and clay, X_1C
6860±130	8.50-8.25 m, silt and clay, X_1C
6030±230	13.50-13.38 m, silt and clay, X_1C (ATY)
5830±160	6.25-6.0 m, silt and clay, X_1C
3070±95	3.25-3.0 m, peat, X_1C

Coding

10,220-6290 yr B.P.	intermediate (2)
6420 -6300 yr B.P.	high (3)
6300-5800 yr B.P.	intermediate (2)
5800-0 yr B.P.	low (1)

Preliminary coding: February 1999
Final coding: 14-03-1999
Coded by BX, GY and SPH

3.12. Chaerhan Salt Lake, Qinghai Province

Chaerhan Salt Lake (Chaerhan in standard Chinese phonetics, given as Qarhan in Zheng et al., 1989; 36.63-37.22°N, 93.72-96.25°E, 2675m above sea level) is a large playa surface in the Chaerhan Basin, which lies in the Chaidamu region of inland Qinghai. The Chaerhan Salt Lake playa has an area of 5856km^2 and is one of largest playa surfaces in the world. Within the playa basin there are nine salt lakes (Dabiele, which is also referred to as Dabieletan or Bieletan; Xiaobiele, which is also referred to as Xiaobieletan; Sheli, which is also referred to as Suli Hu; Xidabuxun; Dabuxun, which is also referred to as Dabsan Hu; Tuanjie; Nanhuobuxun, which is also referred to as Nan Hulsan Hu; Beihuobuxun, which is also referred to as Bei Hulsan Hu; and Xiezuo), with a total area of 460km^2 (Zheng et al., 1989). The elevations of the modern lakes vary from 2675 m a.s.l. (Dabuxun, Beihuobuxun) to 2680 m a.s.l. (Xiezuo). Dabuxun Salt Lake, with area of 184km^2, lies in the western part of the Chaerhan basin and is the largest of the nine lakes. The nine salt lakes have water depths between 1m and 0.2m. Dabuxun Salt Lake has a maximum depth of 0.39m. The salt contents of the lakes varies between 164.81-359.50 g/L and the pH values are between 5.4-7.85. Dabuxun Salt Lake has a salt content of 318.56 g/L and a pH of 7.35. There is no outflow from the Chaerhan Basin, but 7 intermittent streams and 6 permanent streams bring water into the basin (Zheng et al., 1989). The salt lakes are supported by groundwater (1% of total input to the lakes) and surface runoff (99%) derived from precipitation over the catchment and snow/ice meltwater from the surrounding high mountains (Zheng et al., 1989). The Chaerhan Basin originated through faulted structures in the Mesozoic. The underlying bedrock is formed from Proterozoic metamorphics, Palaeozoic sandstone and limestone, and Mesozoic granite. Lacustrine deposition with the Chaerhan Basin has been continuous since the late Cenozoic. There has been slow subsidence of the Chaerhan Basin throughout the Quaternary. As a result, Quaternary lacustrine deposits have a thickness of 2500m (Zheng et al., 1989). However, there is no evidence for abrupt tectonic activity that might have disrupted the drainage network within the basin or otherwise affected the record of changes in relative lake depth (Zheng et al., 1989). The climate in the catchment is cold (0.1°C annual mean temperature) and extremely arid with 28-40 mm total annual precipitation but ca 3000 mm total annual evaporation (Du and Kong, 1983). The Chaerhan Basin is characterised by sparse desert shrubs and forbs, and dominated by Chenopodiaceae, *Ephedra* and *Tamarix* (Du and Kong, 1983).

A number of deep cores have been taken from the Chaerhan Basin in order to investigate salt resources (Zheng et al., 1989). The stratigraphy of these cores indicates that there was a freshwater lake in the basin between 2,470,000 yr B.P. (by palaeomagnetic chronology) and 25,000 yr B.P. (by ^{14}C dating). The lake became salt after 25,000 yr B.P. (by ^{14}C dating) and subsequently (ca 8000 yr B.P.) became a playa. Zheng et al. (1989) showed that there were at least three salt layers (S1, S2 and S3) deposited across most parts of the Chaerhan Basin, although two additional salt layers (S1, S2-1, S2-2, S3-1 and S3-2) are recognised in the central part of the basin, including in Dabuxun Salt Lake. Intercalated with these salt layers are 3-5 detrital units, indicating intervals of relatively freshwater conditions during the last 25,000 yr B.P. (Zheng et al., 1989). The area of the oldest salt layer (S1) is 3086km^2, of the second salt layer (S2) 2300km^2, and of the youngest (S3) 5856km^2.

An additional 910-m long core (CK-6) from the centre of the Chaerhan Basin provides a sedimentary record back to ca 730,000 yr B.P. (Huang and Chen, 1990). The chronology is based on 3 radiocarbon dates from the uppermost 55m, 15 uranium/thorium dates between 392-57m, and palaeomagnetic measurements (the Brunhes/Matuyama boundary occurs at 850m). Only a rudimentary description of the core stratigraphy has been published. According to this description, the uppermost 55m of the core (i.e. covering the last 32,000 years) consists mainly of halite deposits and represents a relatively low phase in the lake history. The underlying deposits are lacustrine or freshwater swamp deposits, and indicate a long period (32,000 to 790,000 yr B.P.) when the climate was significantly less arid than in the recent past.

The more recent history of the Chaerhan Basin can be reconstructed from geomorphic evidence and from a number of shallow cores from the basin floor.

A lacustrine shell-ridge (96.35°E, 36.50°N, 2704 m a.s.l.) is located ca 20km to the east of the modern playa margin. The ridge outcrop is 400m long, 150m wide and +3m high. The ridge contains abundant mollusc shells, including specimens of the bivalves *Corbicula largillierti* and *C. fluminea*, and also *Cyraulus albus* and *Lymnaca* sp. The ridge also contains abundant ostracodes, including *Candona neglecta, Candona* sp., *Candoniella albicans, C.* sp., *Cypridopsis* sp., *Cyclocypris* sp., *Limnocythere dubiosa, L. sanctipartricii, L. binoda, Leucocythere* sp. and *Qinghaicypris* sp. Chen et al. (1990) interpreted these fossil assemblages as reflecting fresh to brackish water conditions. Three samples from 0.05-0.25m, 1.1-1.2m and 1.7-1.8m in the ridge profile were radiocarbon-dated to 28,650±670, 35,100±900 and 38,600±680 yr B.P. respectively. Thus, the presence of freshwater fossils in this lacustrine ridge indicates that there was a large freshwater lake, extending to 20km east of the modern playa, between ca 38,000-28,000 yr B.P. The basal elevation of the ridge is 2701 m a.s.l., indicating that the level of this freshwater lake was ca +29m higher than the level of Dabuxun, the lowest of the modern salt lakes in the basin.

A more detailed lithological, geochemical and pollen record of lake status changes in the basin is provided by a 101m-long core (Core CK2022) taken from the playa surface to the east of Sheli Lake and thus between Sheli Lake and Dabiele Lake (Huang et al. 1980; Du and Kong et al., 1983). There are five salt layers in this core. The chronology is provided by 12 radiocarbon dates. There are three other radiocarbon-dated cores (Core 1308, CK826, CK659), but the stratigraphy of these cores has not been described in detail and it is not possible to correlate these cores with CK2022. The radiocarbon dates from these cores are listed below, but the dates are not used to erect a chronology.

The basal sediment in Core CK2022 (101.0-84.0m) is dark-coloured lacustrine muddy clay. The detrital nature of the sediments indicates deep water. The aquatic assemblage is characterized by abundant *Typha* (14.8%) and *Pediastrum*. Two samples from 84.50-85.00m and 96.36-96.76m were radiocarbon-dated to 26,400±700 and 31,900±2000 yr B.P. Extrapolation of the sedimentation rate between these two dates (0.218 cm/yr) suggests this deepwater deposit was already forming by 33,860 yr B.P. and continued to form until 26,400 yr B.P. This phase of deep water deposition appears to be correlated with the formation of the lacustrine shell-ridge between ca 38,000-28,000 yr B.P.

The overlying unit (84.00-77.92m) is yellow lacustrine clay. *Pediastrum* disappears from the aquatic record and is replaced by *Zygnema*. There is a decrease in the abundance of *Typha* (1.1%). The less-organic and finer-grained nature of the deposits suggest that the lake became somewhat deeper. The decreased abundance of *Typha* is consistent with this interpretation. However, Du and Kong (1983) interpret the shift from dark-coloured muddy clay to yellow clay as indicating shallowing of the lake. They base their argument partly on the terrestrial pollen record, which appears to show more arid conditions during the deposition of the yellow clay, and on the shift from *Pediastrum* (which they interpret as a form tolerant of fresh to slightly saline, still water conditions and water depth < 15m) to *Zygnema* (which they interpret as a form tolerant of shallow, still and fresh water conditions). The ecology of the green algae is not well known, and both *Pediastrum* and *Zygnema* appear to grow in a wide variety of habitats, including lakes that are much deeper than 15m. We do not consider the shift in the algal assemblage to be diagnostic, and rely on the change in sedimentology and in the abundance of *Typha* to suggest that the transition to yellow clay marks a deepening of the lake. The yellow clay unit is dated to between 26,400-25,550 yr B.P. The upper boundary is interpolated by sedimentation rate (1.276 cm/yr) between two radiocarbon dates. The increased sedimentation rate from 0.218 to 1.276 cm/yr is consistent with shallower water.

The overlying unit (77.92-70.44m) is variously described as gypsum-bearing silt (Huang et al., 1980) and sandy clay (Du and Kong, 1983). However, according to the diagram in Huang et al. (1980), the unit contains no evaporite minerals. The coarser nature of the sediment suggests decreased water depth. The slight increase in *Typha* (ca 2%) is consistent with decreased water depth. There is no record of green algae within this unit. The unit was formed between 25,550 and 24,970 yr B.P.

The overlying unit (70.44-61.63m) is gypsum-bearing clay. The finer-grained nature of the sediments suggests increased water depth. The unit contains 75-95% detrital minerals and 5-25% evaporite minerals (including gypsum). The increase in evaporitic minerals suggests an increase in water salinity, which does not appear to be consistent with increased water depth. There is no change in the abundance of *Typha*. The significance of the presence of *Pediastrum* and the absence of *Zygnema* is not clear, although Du and Kong (1983) suggest it indicates increased water depth. Given the fact that the increase in the amount of evaporites is small, we suggest that the lake probably became deeper but that any increase in water depth was slight. Two samples from 65.89-66.42m and 68.14-68.54m were radiocarbon-dated to 24,400±510 and 24,800±470 yr B.P. respectively, suggesting deeper water lake phase occurred between 24,970-23,860 yr B.P.

The overlying unit (61.63-39.70m) is a silty gypsum-bearing halite, equivalent to the S1 salt layer defined by Zheng et al. (1989). The unit contains <10% detritus minerals and >90% evaporite minerals. The presence of halite indicates a significant shallowing of the lake. *Typha* is present initially, but disappears from the record thereafter. Both *Zygnema* and *Pediastrum* are present, though *Pediastrum* is less abundant than in the underlying unit. Du and Kong (1983) interpret this evidence as indicating shallower water. The upper boundary is estimated to 21,170 yr B.P. by interpolation of sedimentation rate (0.8855 yr/cm) between the radiocarbon date of 24,400 yr B.P. from the underlying unit and of 20,600 yr B.P. from the overlying unit.

The overlying unit (39.70-34.72m) is gypsum-bearing clay. The unit contains 90% detritus minerals and 10% evaporite minerals. The significant increase in the detritic mineral content, and the shift from halite to gypsum dominance in the evaporitic component of the sediments, indicate a return to deeper water conditions. The unit contains no aquatics and no green algae. A sample from 35.52-34.87m was radiocarbon-dated to 20,600±410 yr B.P., suggesting this unit was formed between 21,170-20,600 yr B.P.

The overlying unit (34.72-22.16m) is a clayey gypsum-bearing halite, equivalent to the S2-1 salt layer defined by Zheng et al. (1989). The unit contains 25-30% detritus minerals and 70-75% evaporite minerals. The higher detrital content, and correspondingly lower evaporitic content, indicates that the lake was somewhat fresher than during the deposition of the first salt layer (S1). This is consistent with the fact that the S2 salt deposits cover a smaller area ($2300km^2$) than the S1 deposits ($3086km^2$). *Typha* is present within the unit (ca 1%) but there is no record of green algae. A sample from 22.16m was radiocarbon-dated to 16,000 yr B.P., suggesting this unit was deposited between 21,170-16,000 yr B.P.

The overlying unit (22.16-19.15m) is gypsum- and halite-bearing clay. The unit contains 70% detritus minerals and 30% evaporite minerals. The finer-grained nature of the sediments, the reduction in the amount of evaporitic minerals, and the increased importance of gypsum relative to halite, suggest the lake became less saline and deeper. *Typha* is present within the unit (ca 1%) but there is no record of green algae. A sample from 21.19-21.39m was radiocarbon-dated to 15,700±340 yr B.P. Using the sedimentation rate (0.290 mm/yr) between this date and the 16,000 yr B.P. date from the lower boundary of the unit, the upper boundary of this unit is dated to 14,960 yr B.P.

The overlying unit (19.15-13.87m) is a clayey gypsum-bearing halite, equivalent to the S2-2 salt layer as defined by Zheng et al. (1989). The unit contains <10% detritus minerals and >90% evaporite minerals. The higher evaporitic content, and correspondingly lower detrital content, of this unit suggests that the lake was more saline than during the formation of the S2-1 unit. *Typha* is present within the unit (ca 1%) but there is no record of green algae. A sample from 13.87 was radiocarbon-dated to 9300± yr B.P., suggesting this salt layer was deposited between 14,960-9300 yr B.P.

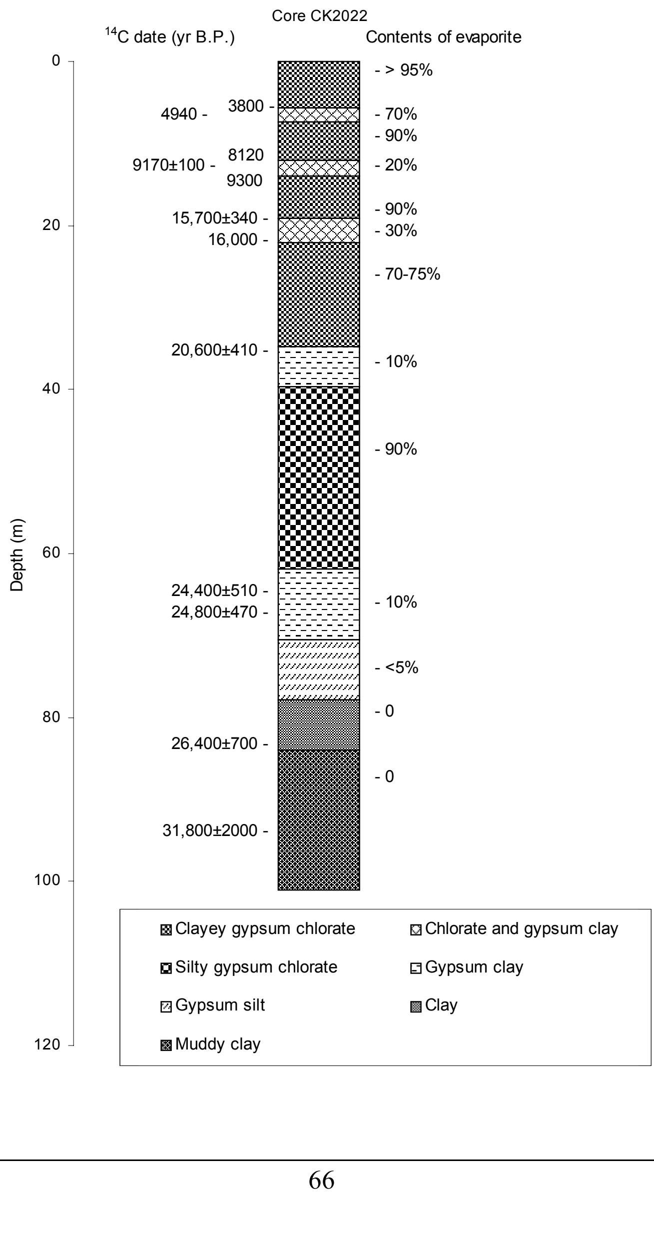
The overlying unit (13.87-12.03m) is gypsum- and halite-bearing clay. The unit contains 80% detrital minerals and 20% evaporite minerals. The increase in the detritic content, and the increase in the relative importance of gypsum compared to halite in the evaporite component, indicate that the lake was less saline than before and suggest it became deeper. *Typha* is present within the unit (ca 1%) but there is no record of green algae. Samples from 13.38-13.78m and 12.03m were radiocarbon-dated to 9170±100 and 8120 yr B.P. respectively, suggesting this unit was formed between 9300-8120 yr B.P.

The overlying unit (12.03-7.31m) is a clayey gypsum-bearing halite, equivalent to the S3-1 salt layer defined by Zheng et al. (1989). The unit contains <10% detrital minerals and >90% evaporite minerals. The increase in evaporite content, and in the relative importance of halite compared to gypsum, suggests the lake became more saline and shallower. There is no record of aquatics or green algae within the unit. A sample from 7.31m was radiocarbon-dated to 4940 yr B.P., suggesting this salt layer was deposited between 8120-4940 yr B.P.

The overlying unit (7.31-5.67m) is gypsum- and halite-bearing clay. The unit contains 30% detrital minerals and 70% evaporite minerals. The decrease in evaporite minerals, and the increased importance of gypsum compared to halite, suggests the lake became somewhat less saline and deeper than formerly. There is no record of aquatics or green algae within the unit. A sample from 5.67m was radiocarbon-dated to 3800 yr B.P., suggesting this unit formed between 4940-3800 yr B.P.

The uppermost unit (5.67-0m) is a clayey gypsum-bearing halite, equivalent to the S3-2 salt layer defined by Zheng et al. (1989). The unit contains >95 % evaporite minerals, suggesting extremely saline and shallow water conditions between 3800-0 yr B.P. The area of the salt deposit (5856km^2) is larger than any of the other salt units found in Chaerhan, consistent with the interpretation that the present represents the most arid phase in the history of this basin. There is no record of the presence of green algae. However, the presence and relative abundance of *Typha* (4-5%) within this unit is consistent with shallow-water conditions.

In the status coding, extremely low (1) is indicated by clayey gypsum-bearing halite with >95% evaporites (unit S3); very low (2) by clayey/silty gypsum-bearing halite >90% evaporites; low (3) by clayey gypsum-bearing halite or gypsum-halite clay with 70-80% evaporites; moderately low (4) by gypsum- and halite-bearing clay with 20-30% evaporites; intermediate (5) by gypsum-bearing clay with <25% evaporites and no halite; moderately high (6) by silt or sandy clay containing no evaporites; high (7) by organic muddy clay with no evaporites and high amounts of *Typha*; and very high (8) by yellow lacustrine clay with no evaporites and low amounts of *Typha*.



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Radiocarbon dates

38,600±680	1.70-1.80m, carbonate, shell-ridge
35,100±900	1.10-1.20m, carbonate, shell-ridge
33,800±3000	14.15-14.65m, salt-bearing yellow clay, Core CK826
32,200±1800	54.9m, halite deposit, Core CK-6
31,800±2000	96.36-96.76m, black clay, Core CK2022
29,700±500	44.17-44.52m, dark grey clay, Core 1308
28,650±670	0.05-0.25m, carbonate, shell-ridge
28,200±900	52.3m, halite deposit, Core CK-6
27,600±1100	16.30-16.70m, salt-bearing clay, CK659
26,400±700	84.50-85.00m, clay, Core CK2022
24,800±470	68.14-68.54m, gypsum-bearing clay, Core CK2022
24,800±900	40.8m, halite deposit, Core CK-6
24,400±510	65.89-66.42m, gypsum-bearing clay, Core CK2022
21,200±1050	7.72-8.22m, salt-bearing black muddy clay, Core CK826
21,200±210	27.02-27.42m, salt-bearing black muddy clay, Core 1308
20,600±410	35.52-34.87m, salt-bearing clay, Core CK2022
18,100±500	25.88-26.28m, salt-bearing gypsum clay, Core 1308
16,000±	22.16m, salt-bearing clay, Core CK2022
15,700±340	21.19-21.39m, salt-bearing clay, Core CK2022
14,900±100	14.89-15.19m, salt-bearing clay, Core 1308
13,000±400	5.66-6.04m, salt-bearing clay, Core CK826
9310±310	2.5-2.9m, salt-bearing muddy clay, Core 1308
9300±	13.87m, salt-bearing clay, Core CK2022
9170±100	13.38-13.78m, salt-bearing clay, Core CK2022
8850±210	1.30-1.70m, salt-bearing black muddy clay, Core CK826
8120±	12.03m, salt-bearing clay, Core CK2022
4940±	7.31m, salt-bearing clay, Core CK2022
3800±	5.67m, salt-bearing clay, Core CK2022

(The samples were dated from ^{14}C Lab of Institute of Salt Lake, Chinese Academy of Science.)

U/Th dates

57,500±8300 yr B.P.	57m, Core CK-6
74,800±9800 yr B.P.	92.0m, Core CK-6
82,300±10900 yr B.P.	99.2m, Core CK-6
104,000±9400 yr B.P.	151.9m, Core CK-6
119,500±11900 yr B.P.	169.6m, Core CK-6
191,900±24500 yr B.P.	206.7m, Core CK-6
192,900±34000 yr B.P.	249.7m, Core CK-6
204,000±38800 yr B.P.	260.9m, Core CK-6
265,000±43000 yr B.P.	271.2m, Core CK-6
254,200±45000 yr B.P.	283.4m, Core CK-6
277,700±73000 yr B.P.	297.7m, Core CK-6
257,600±60000 yr B.P.	323.0m, Core CK-6
299,000±92000 yr B.P.	354.0m, Core CK-6
341,000+213,000-83,000 yr B.P.	378.2m, Core CK-6
336,000+500,000-77,000 yr B.P.	392.2m, Core CK-6

Coding

38,000-26,400 yr B.P.	high (7)
26,400-25,550 yr B.P.	very high (8)
25,550 -24,970 yr B.P.	moderately high (6)
24,970-23,860 yr B.P.	intermediate (5)
23,860-21,170 yr B.P.	very low (2)
21,170-20,600 yr B.P.	intermediate (5)
20,600-16,000 yr B.P.	low (3)
16,000-14,960 yr B.P.	moderately low (4)
14,960-9300 yr B.P.	very low (2)
9300-8120 yr B.P.	moderately low (4)
8120-4940 yr B.P.	very low (2)
4940-3800 yr B.P.	low (3)
3800-0 yr B.P.	extremely low (1)

Preliminary coding: 29-01-1999
Second coding: 18-9-2000
Final coding: 18-11-2000
Coded by GY and SPH

3.13. Dachaidan-Xiaochaidan Salt Lakes, Qinghai Province

Dachaidan-Xiaochaidan Salt Lakes (Dachaidan and Xiaochaidan in standard Chinese phonetics, given as Da Qaidam and Xiao Qaidam by Zheng et al., 1989) are two independent, closed salt lakes in the Chaidamu region of inland Qinghai. The lakes are ca 30km distant from one another. However, the two lakes were a united as a single freshwater lake before 30,000 yr B.P. (Zheng et al., 1989) and we therefore treat them as a single lake for the purposes of coding the lake status record. Dachaidan Salt Lake (37.83°N, 95.23°E, 3110m above sea level) has a water area which varies seasonally from 35.9km^2 in the summer flood season to 22.92km^2 in the winter dry season. The playa area (i.e. the low-lying part of the basin floor adjacent to the lake and subject to salt efflorescence) is 240km^2 (Zheng et al., 1989). This salt lake has a maximum depth of 0.7m and a mean depth of 0.34m during the summer flood season. The salt content of the water varies seasonally between 103.6-387 g/L, with a maximum in the winter season and a minimum in the summer season (Zheng et al., 1989). The Dachaidan Salt Lake has no outflow, but there are four seasonally-intermittent streams (Yuka, Wenquangou, Baligou and Datouyanggou Rivers) flowing into the lake. The Xiaochaidan Salt Lake (37.17°N, 95.50°E, 3118m a.s.l.) has a water area of 69km^2 and a playa area of 150km^2. A single river (Tataleng) feeds the lake. The salt content of the water is 325-339.1 g/L. The combined catchment of Dachaidan-Xiaochaidan has an area of 3,100 km^2. The Dachaidan-Xiaochaidan is supported by surface runoff from precipitation in the catchment, by snow/ice meltwater from high mountains, and by some warm-springs associated with underground tectonic faulting (Zheng et al., 1989). The Dachaidan-Xiaochaidan basin originated through faulting in the Mesozoic. The underlying bedrock consists of Proterozoic metamorphics, and Mesozoic sandstones and granite. The climate in the catchment is cold (0°C annual mean temperature) and extremely arid, with a total annual precipitation of 88.4mm and a total annual evaporation of 2080mm (Zheng et al., 1989).

Many deep cores (maximum depth up to 100.88m) have been taken from the Dachaidan-Xiaochaidan Basin in order to investigate salt-mineral resources (Zheng et al., 1989). The stratigraphy of these cores shows a change from alluvial-fluvial sediments in the Lower and Middle Pleistocene to lacustrine deposition in the Upper Pleistocene (Zheng et al., 1989). The basal lacustrine deposits are distributed continuously over large areas within the Dachaidan-Xiaochaidan basin, indicating that the modern Dachaidan and Xiaochaidan Lakes are relicts of a former large palaeolake (Zheng et al., 1989). This large palaeolake occupied the Daichaidan-Xioachaidan basin sometime prior to 30,000 yr B.P. Subsequently the lake shrank because of long-term aridification and the consequent decrease in the water supply to the catchment. The Daichaidan and Xiaochaidan Lakes became separated at this stage. Fluvial-alluvial deposits overlying the basal lacustrine deposits in the region between the two lakes testify to this separation sometime during the Late Pleistocene.

Since the Dachaidan and Xiaochaidan Lakes have been separated, they have undergone significant changes in depth and area. Intervals when the lake was high and fresh (indicated by primarily detrital deposits which do not contain halite salts) alternated with more arid periods when the lake was lower and salt-rich deposits were formed. Zheng et al. (1989) reported the existence of 4 halite-salt layers in the bottom-floor deposits of Dachaidan Salt Lake. The lowermost of these salt-rich units are underlain by detrital clays, and indicate an initial phase of freshwater conditions. The first salt-rich layer is 3-

8m thick and covers an area of ca 30km^2. The second salt-rich layer is 3-4m thick and covers an area of ca 94km^2. The third salt-rich layer is 3-4m thick and covers an area of ca 56km^2. The fourth salt-rich layer is 6-8m thick and covers an area of 109km^2.

A more detailed record of lake-depth changes in the Dachaidan Salt Lake during the Late Quaternary is provided by an 11m-long core (Core CK3) taken from lakeside playa (Huang et al. 1980). The same four salt layers recognised by Zheng et al. (1989) were found in Core CK3. A second core (CK2022) was taken from Bieletang sub-basin. This core shows the same basic stratigraphy as CK3, but it has some additional halite-rich layers because it lies near the basin margin and at higher elevation and thus was more susceptible to drying out. Although there are a number of cores from Xiaochaidan (Huang et al., 1980; Zheng et al., 1989), the stratigraphic record of these cores is not published. Therefore our reconstructions of changes in water depth and salinity in the Dachaidan-Xiaochaidan basin are based on changes in lithology and geochemistry recorded in the Dachaidan Core CK3. The chronology is based on 3 radiocarbon dates from that core. Confirmation of the chronology of changes in lake depth and/or area is provided by reference to the radiocarbon dating of similar units in Core CK2022. The reconstructed changes are concordant with the gross changes documented by Zheng et al. (1989).

The basal unit (11.0-9.84m) in Core CK3 is yellow lacustrine clay, suggesting deep water conditions. The unit does not contain evaporite minerals, consistent with fresh water and deep lake conditions. A sample from 9.84-9.40m was radiocarbon-dated to 21,000±1060 yr B.P. Using the sedimentation rate (0.0403 cm/yr) between this date and a date of 14,300±460 yr B.P. from a depth of 6.77-7.06m, this unit was formed between 24,800-21,950 yr B.P.

The overlying unit (9.84-9.02m) in Core CK3 is black lacustrine muddy clay. The more organic nature of the sediment suggest that the lake became shallower. The presence of gypsum and borate in low abundance within this unit indicates somewhat increased salinity and is consistent with shallowing. This unit is dated to between 21,950-19,510 yr B.P.

The basal deposits in Core CK2022 consist of black muds, that are presumed to represent deposition in the united Dachaidan-Xiaochaidan megalake. Two samples from these units have been radiocarbon-dated to 31,800±2000 and 26,400±700 yr B.P. respectively. Fluvial deposits (clays and sand muds) testify to the separation of the sub-basins. The overlying units consist of a sequence of evaporite-free clay, halite-rich clay and evaporite-free clay. The two evaporite-free clays, which are believed to be equivalent to the basal clays in Core CK3, have been radiocarbon-dated to 24,800±470/24,440±510 yr B.P. and 20,600±410 yr B.P. respectively. Thus the record from Core CK2022 confirms that conditions in the Daichaidan-Xiaochiadan region were sufficiently wet to maintain deepwater lakes during the glacial.

The overlying unit (9.02-7.91m) in Core CK3 is a clayey borate-gypsum deposit. The unit contains 75-90% evaporite minerals, mostly gypsum. There is no halite present. The increased abundance of evaporatic minerals indicates that the lake became more saline and is consistent with shallower conditions. However, despite the abundance of evaporatic minerals, the absence of halite suggests that the lake was not at saline as subsequently. This unit is dated to between 19,510-16,800 yr B.P.

The overlying unit (7.91-6.8m) in Core CK3 is clayey gypsum-halite deposit. The unit contains 60% evaporite minerals, of which 40% (i.e. 2/3rds of the evaporites) is halite. The increase in the abundance of halite indicates increased salinity and implies decreased water depth. A sample from near the top of this unit (6.77-7.06m) was radiocarbon-dated to 14,300±460 yr B.P., suggesting the unit was deposited between 16,800-14,300 yr B.P. Zheng et al. (1989) estimated that the area of the halite-salt lake was ca 30km^2. A similar halite-rich layer is found in Core CK2022, but cessation of halite-deposition in the Bieletang sub-basin appears to have ceased somewhat earlier than at CK3.

The overlying sediment (6.8-5.77m) in Core CK3 is halite-gypsum clay. The unit contains 40% evaporite minerals of which < 5% is halite. The decrease in the abundance of halite is consistent with freshening. Interpolation of the sedimentation rate (0.0563 cm/yr) between the date from the underlying unit and a date of 7630±140 yr B.P. from a depth of 2.94-3.39m suggests that this unit was formed between 14,300-12,300 yr B.P. This fresher water deposit is presumed to be equivalent to a clay deposit found in Core CK2022, dated to 15,700±340 yr B.P. (i.e. somewhat earlier than in the Dachaidan sub-basin).

The overlying unit (5.77-4.27m) in Core CK3 is a clayey/sandy halite deposit. The unit contains 90% evaporite minerals, of which 75% is halite. The marked increase in the abundance of halite indicates that the lake water became rather saline, and implies that the lake became shallow. The increased coarseness of the detrital component of the unit is consistent with shallowing. This unit is dated to between 12,300-9600 yr B.P. This unit can be correlated with the second salt unit identified by Zheng et al. (1989). Zheng et al. (1989) estimated that the halite-salt lake had an area of 94km^2. The fact that the salt area is larger than the area during the first phase of halite-salt deposition (30km^2) suggests that conditions in the basin were more saline than during the depositon of the first halite-salt deposits. Again, a similar halite-rich layer is found in Core CK2022.

The overlying unit (4.27-2.94m) in Core CK3 is halite-gypsum clay. The unit contains 90% detritus minerals and 10% evaporite minerals. The change in lithology indicates increased water depth. The abundance of detrital minerals and the significant decrease in evaporite minerals is consistent with increased water depth and relatively freshwater conditions. A sample from the top of the unit (2.94-3.39m) was radiocarbon-dated to 7630±140 yr B.P., suggesting the unit was formed between 9600-7080 yr B.P. The equivalent clay unit in Core CK2022 is radiocarbon-dated to 9170±100 yr B.P., and thus provides confirmation of the timing of this wetter interval in the Dachaidan-Xiaochaidan region.

The overlying unit (2.94-1.49m) is a clayey/sandy borate-gypsum-halite deposit. The unit contains 80-95% evaporite minerals, of which 70-75% is halite. The increased coarseness of the sediments indicates shallower water conditions. The significant increase in evaporite minerals is consistent with increased salinity and shallowing. Interpolation of the sedimentation rate (0.0415 cm/yr) between the radiocarbon date from the underlying unit and the core top, which is assumed to be modern, suggests this unit was formed between 7080-3590 yr B.P. This unit is equivalent to the third salt-bearing unit identified by Zheng et al. (1989). Zheng et al. (1989) estimated the halite-salt lake had an area of 56km^2. The small area of the salt deposits compared with the area of the early Holocene salt deposits (94km^2) indicates that this phase was not as saline as during the earlier interval.

The overlying unit (1.49-0.8m) is a clayey/sandy gypsum-halite deposit. The unit contains 25% detritus minerals and 75% evaporite minerals, of which 50% is halite. The increase in detrital minerals suggests water depth increased slightly and the lake became less saline. However, the abundance of halite minerals indicates that the lake was not fresh. This unit is dated to between 3590-1930 yr B.P.

The uppermost unit (0.8-0m) is a gypsum-halite deposit. The unit contains > 95% evaporite minerals, almost all of which is halite. The increase in the abundance of evaporite minerals and the dominance of halite indicates extremely saline and shallow water conditions after 1930 yr B.P. This unit is equivalent to the fourth saline unit identified by Zheng et al. (1989). Zheng et al. (1989) estimated this salt lake had a maximum area of 109km^2. The expanse of the halite-salt deposits is consistent with the interpretation of extremely saline conditions.

In the status coding, extremely low (1) is indicated by extensive (109km^2) halite-salt deposits in the basin and gypsum-halite deposits with >95% evaporites in Core CK3; very low (2) by less extensive (94km^2) halite-salt deposits in the basin and clayey/sandy gypsum-halite deposits with > 90% evaporites in Core CK3; low (3) by moderately confined (56km^2) halite-salt deposits in the basin and clayey/sandy gypsum-halite deposits with 80-95% evaporites in Core CK3; moderately low (4) by clayey gypsum-halite deposits with 75% evaporite minerals in Core CK3; intermediate (5) by confined (30km^2) halite-salt deposits and by clayey gypsum-halite deposits with 60% evaporite minerals, of which 40% is halite, in Core CK3; moderately high (6) by clayey halite-gypsum deposits with 60% evaporite minerals, of which only 5% is halite, in Core CK3; high (7) by clayey halite-gypsum deposits with only 10% evaporite minerals and less than 5% halite in Core CK3; very high (8) by clayey borate-gypsum deposits with 75-90% evaporites but not containing halite in Core CK3; extremely high (9) by organic clays with <10% evaporite minerals in Core CK3; and maximally high (10) by evaporite-free clays in Core CK3.

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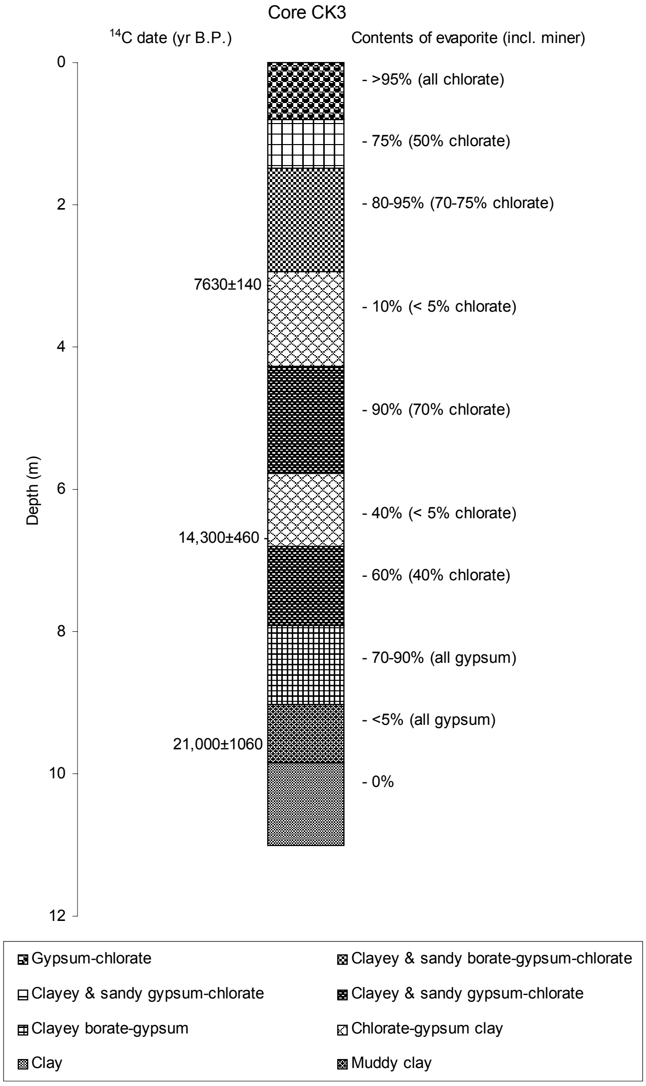
Radiocarbon dates

7630±140 yr B.P.	2.94-3.39m, salt-bearing clay, Core CK3
9170±100 yr B.P.	13.38-13.78m, salt-bearing clay, Core CK2022
14,300±460 yr B.P.	6.77-7.06m, salt-bearing clay, Core CK3
15,700±340 yr B.P.	21.19-21.39m, salt-bearing clay, Core CK2022
20,600±410 yr B.P.	34.52-34.87m, salt-bearing clay, Core CK2022
21,000±1060 yr B.P.	9.40-9.84m, clay, Core CK3
24,440±510 yr B.P.	65.89-66.42m, salt-bearing clay, Core CK2022
24,800±470 yr B.P.	68.14-68.54m, salt-bearing clay, Core CK2022
26,800±700 yr B.P.	84.50-85.00m, mud, Core CK2022
31,800±2000 yr B.P.	96.36-96.76m, black mud, Core CK2022

(The samples were dated by the ^{14}C Lab of Institute of Salt Lake, Chinese Academy of Science.)



Lake status

24,800-21,950 yr B.P.	maximally high (10)
21,950-19,510 yr B.P.	extremely high (9)
19,510-16,800 yr B.P.	very high (8)
16,800-14,300 yr B.P.	intermediate (5)
14,300-12,300 yr B.P.	moderately high (6)
12,300-9600 yr B.P.	very low (2)
9600-7080 yr B.P.	high (7)
7080-3590 yr B.P.	low (3)
3590-1930 yr B.P.	moderately low (4)
1930-0 yr B.P.	extremely low (1)

Preliminary coding: 01-03-1999
Second coding: 16-09-2000
Final coding: 16-11-2000
Coded by GY and SPH

3.14. Gounongcuo, Qinghai Province

Gounongcuo (34.35°N, 92.2°E, 4670m above sea level; coordinates given in Wang et al., 1987; Li et al, 1994; and Li et al., 1995 are different; we used the one given in the latest paper) is a salt lake in a closed basin in the Kekexili Region, northern Tibetan Plateau. The lake overflows in the summer season but there is no discharge in the winter season. The lake has an area of 2.9 km^2 (Wang et al., 1987). The lake water is supplied mainly by runoff from the basin (Li et al., 1995). The climate is cold (0°C mean annual temperature) and dry (200mm total annual precipitation but 2000mm annual evaporation) (Li et al., 1995). The Kekexili Region is characterised by alpine meadows, steppe and desert vegetation, with abundant Chenopodiaceae, *Ephedra* and *Artemisia* (Shan et al., 1996).

A 7.25m-long core (Core KX-1) from the lake flat, 20cm above the current lake level, provides a sedimentary record back to before 20,000 yr B.P. (Li et al., 1994; 1995). Changes in water depth and salinity are reconstructed on the basis of changes in lithology, ostracode assemblages, and aquatic macrofossil and pollen assemblages (Li et al., 1994; Shan et al., 1996; Li, 1996). The chronology is based on three radiocarbon dates on bulk sediments from the core for the period between the last glacial maximum and the late glacial (Li et al., 1995). The Holocene chronology is based on sedimentation rate from the core (Shan et al., 1996), and correlation with the standard pollen chronology from western regions of China (Kong and Du, 1991; Shi et al., 1992).

The basal sediments (below 535cm) in the core are brown and reddish sandy clay with muddy fine pebbles. The coarse mineral deposits, without ostracodes or pollen, may represent alluvial debris, indicating that the lake was not formed until after ca 20,000 yr B.P. (Li et al., 1994).

The overlying sediment (513-535cm) is lacustrine grey silty clay with clay silt, suggesting a moderately deep lake. The ostracode assemblage is characterised by the occurrence of *Limnocythere dubiosa*, a species characteristic of fresh to slightly saline water, consistent with a moderately deep freshwater lake (Li et al., 1995). There are only a few grains of aquatic pollen (specific pollen taxa are not given in Shan et al., 1996) in the unit, consistent with this interpretation. A sample from near the bottom of this unit (530cm) is radiocarbon-dated to 19,210±480 yr B.P. The upper boundary is dated to ca 17,200 yr B.P. by interpolation between radiocarbon dates.

The overlying sediment (413-513cm) is lacustrine laminated, grey-black clay and silty-clay, suggesting increased water depth. The ostracode assemblage is characterised by *Limnocythere dubiosa* with abundant *Ilyocypris biplicata* (a general fresh water species: Li, 1996), consistent with increased water depth. The paucity of aquatic pollen is consistent with this interpretation. The unit is dated to between 17,200 and ca 15,800 yr B.P. by interpolation between radiocarbon dates.

A decrease in water depth after 15,800 yr B.P. is indicated by the occurrence of 6 layers of peat each 0.5-5cm thick, separated by 5 intervals of green clay deposition (293-413cm). Both the peats and the clays contain abundant plant debris, including fruits of *Potamogeton pectinatus* and *P. perfoliatus*, and leaves of *Vallisneria* (Shan et al., 1996). The aquatic pollen assemblage is characterised by abundant *Typha*, with *Pediastrum boryanum* and *Spirogyra*, consistent with shallow water. Two samples from the peat at

382cm and 315cm are radiocarbon-dated to 15,237±461 yr B.P. and 13,035±155 yr B.P. respectively.

The overlying sediment is grey silty clay (226-293cm), suggesting a return to deeper water after 13,000 yr B.P. The presence of *Eucypris inflata*, an ostracode species that tolerates extreme salinity up to 110 ‰ (Li et al., 1995) in the upper part (226-263cm) of this silty clay unit, suggests the lake became saline and water depth decreased after ca 11,500 yr B.P. The absence of aquatic pollen (220-260cm) is consistent with salty, shallow water.

The overlying sediment is grey-green clay (165-226cm), suggesting increased water depth after 10,200 yr B.P. The presence of abundant aquatic pollen (*Potamogeton*) (220-198cm) is consistent with increased water depth. The disappearance of *Eucypris inflata* in the upper part of the unit (165-210cm) reflects decreased salinity, suggesting a further increase in water depth between 9800 and 9200 yr B.P. The decrease in *Potamogeton* (198-137cm) is consistent with this interpretation.

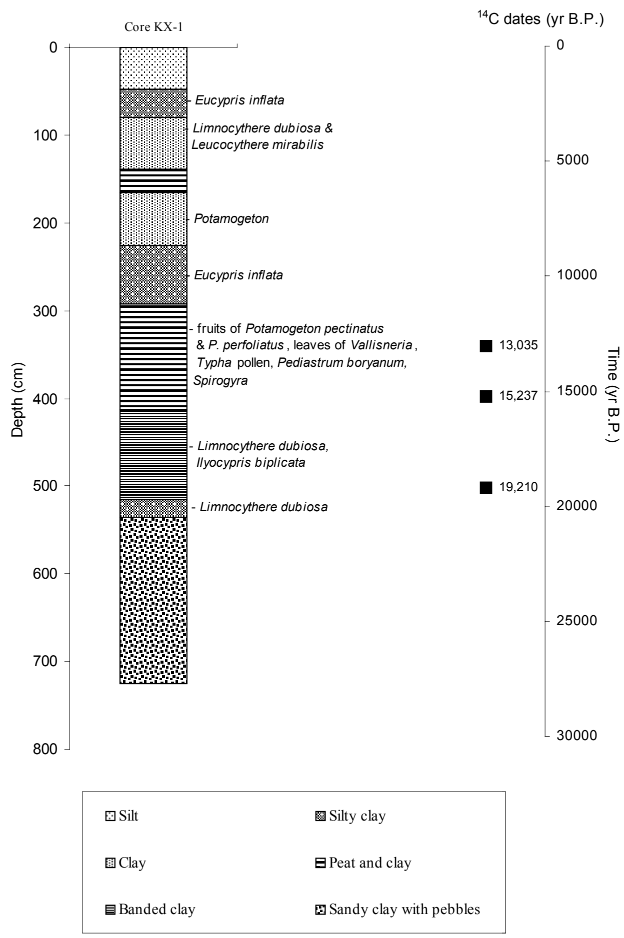
The overlying sediment consists of three layers of peat (165-164cm, 157-159cm and 139-142cm), separated by intervals of thin grey-green clay deposition, suggesting fluctuating but shallow water. The whole unit is dated to between 9200 and 7800 yr B.P.

The overlying sediment is grey-black lacustrine clay (139-80cm), suggesting a return to deep water. The unit is dated to between 7800 and 5500 yr B.P. The ostracode assemblage of *Limnocythere dubiosa* and *Leucocythere mirabilis* (*Leucocythere mirabilis* is a species characteristic of a wide range of salinities between 0.48g/L and 256.73g/L, but its presence with *Limnocythere dubiosa* probably reflects a saline condition based on investigations of modern lakes in Tibet: Li, 1996), and increases in ostracode abundance are consistent with increased depth after ca 8000 yr B.P. (Li, 1996).

The overlying sediment is grey silty clay (80-48cm), suggesting a decrease in water depth. The increase in *Eucypris inflata* within this unit reflects increased salinity and reduced depth between 5500 and 3000 yr B.P. The decrease in ostracode abundance after ca 5500 yr B.P. (Li, 1996) is also consistent with shallowing.

The uppermost sediment (above 48cm) is ginger-coloured silt, fine sand and coarse sand with small pebbles. This is probably a nearshore deposit, suggesting the lake area was reduced and water depth decreased after ca 3000 yr B.P.

In the status coding, low (1) is indicated by lacustrine nearshore deposits of silt, fine and coarse sand; moderately low (2) by lacustrine silty clay with an ostracode assemblage including extreme saline species of *Eucypris inflata,* and aquatic pollen assemblage including *Typha*; intermediate (3) by peat between lacustrine deposits; moderately high (4) by silty clay with an ostracode assemblage including *Limnocythere dubiosa* but not *Eucypris inflata*, and an aquatic pollen assemblage including *Potamogeton*; high (5) by lacustrine clay with laminae and *Ilyocypris biplicata*. The basal alluvial deposit indicates the basin was dry (0) before ca 19,600 yr B.P.



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Radiocarbon dates

13,035±155	peat, 315 cm deep in Core KX-1
15,237±461	peat, 382 cm deep in Core KX-1
19,210±480	silty clay, 530 cm deep in Core KX-1

(Lab numbers are not given; the samples were dated in Archaeology Institute of Chinese Academy of Society)

Coding

pre 19,600 yr B.P.	dry (0)
19,600-17,200 yr B.P.	moderately high (4)
17,200-15,800 yr B.P.	high (5)
15,800-13,000 yr B.P.	intermediate (3)
13,000-11,500 yr B.P.	moderately high (4)
11,500-10,200 yr B.P.	moderately low (2)
10,200-9800 yr B.P.	moderately high (4)
9800-9200 yr B.P.	moderately low (2)
9200-7800 yr B.P.	intermediate (3)
7800-5800 yr B.P.	moderately high (4)
5800-5000 yr B.P.	moderately low (2)
5000-0 yr B.P.	low (1)

Preliminary coding: 03-11-1998
Second coding: 20 -11-1998
Third coding: 1-12-1998
Final coding: November 1999
Coded by GY and SPH

3.15. Wulanwula Lake, Qinghai Province

Wulanwula Lake (given as Ulan U1 Lake by Shan et al., 1996; 34.8°N, 90.5°E, 4854 m above sea level) is a large semi-salt lake in a closed basin in the Kekexili Region, northern Tibetan Plateau. The basin originated as a fault depression (Hu, 1995). The lake has an area of 544.5 km^2. The water depth is unclear, but the maximum depth is at least 6.9m since a core from the lake center was taken in this depth of water (Hu, 1995). The lake water is supplied mainly by runoff from the basin (Li et al., 1995). Climate in the catchment is cold (–5.9°C mean annual temperature) and dry (370 mm total annual precipitation but more than 2000mm of annual evaporation) (Li, 1996). The Kekexili Region is characterised by alpine meadows, steppe and desert vegetation, including Chenopodiaceae, *Ephedra* and *Artemisia* (Shan et al., 1996).

Evidence of higher lake levels during the last glacial maximum and the late glacial is provided by two beach-rock ridges at ca 150cm and 50cm above modern lake level (Hu, 1995). The stratigraphy of these ridges has been studied in two profiles: Profile III from the +150cm ridge and Profile II from the +50cm ridge. A third profile (Profile I) was taken in the lake flat, ca 20cm above modern lake level. Two cores, a 60cm-long core taken from a depth of 6.9m at the lake center (Hu, 1995) and a second 65cm-long core from a depth of 6.0m near the lake center (Shan et al., 1996), provide additional sedimentary records back to ca 11,500 yr B.P. Changes in water depth are reconstructed on the basis of geomorphological evidence and changes in lithology and aquatic pollen assemblages (Hu, 1995; Shan et al., 1996). The chronology is based on five pre-Holocene radiocarbon dates (Hu 1995; Shan et al., 1996). The Holocene chronology is based on extrapolation of the sedimentation rate from the radiocarbon dated part of the cores (Shan et al., 1996), assuming that the core top is modern, but is consistent with the standard pollen chronology from western regions of China (Kong and Du, 1991; Shi et al., 1992).

The beach-rock ridge, ca 150cm above the modern lake level and distributed intermittently around the lake (Hu, 1995) provides evidence of a high lake stand during the last glacial maximum. The profile (profile III) from this ridge shows pebbles and sands at the base (below ca 20cm) overlain by brown lake mud and clay (ca 20cm thick) which was concreted into lacustrine beach-rock after lake level lowered. A sample from the lacustrine unit is radiocarbon-dated to 18,217±390 yr B.P. This suggests a high lake stand, ca 150cm higher than present before 18,000 yr B.P.

A second beach-rock ridge, ca 50cm above the modern lake level and distributed intermittently around the lake (Hu, 1995) provides evidence for a later period of higher-than-present lake level. The sediment profile (Profile II) taken from this ridge, 160m away from the lake margin, shows a light brown lake mud (ca 20cm thick) concreted into lacustrine beach-rock. A sample from the unit is radiocarbon-dated to 10,997±252 yr B.P. This suggests a lake stand ca 50cm higher than present ca 11,000 yr B.P.

A sediment profile (Profile I) was dug from lake flat, a few of meters away from the lake margin, with the top at 20cm above lake level. The basal sediments (10-30cm) are lacustrine silt unit which shows colour-bedding of grey-blue and brown. The overlying unit (5-10cm) is grey lacustrine muddy silt. The overlying sediment is a 3-5cm thick, ginger-coloured, fine sand interpreted as beach sand. The uppermost unit is a 1-2cm thick salt crust (Hu, 1995). The basal sediment and the overlying lacustrine silt unit suggest a high lake level at least 15cm above modern lake level. A sample from the top of the lacustrine muddy silt is radiocarbon-dated to 12,359±253 yr B.P. Thus this high lake phase must have terminated by ca 12,000 yr B.P., but there is no information about the duration of this interval.

The two cores from the centre of the lake apparently cover the late glacial and Holocene. There is no information about the 60cm long core, except that the basal sediments (50-60cm) are black lacustrine mud (Hu, 1995). This unit has been radiocarbon-dated to 11,313±212 yr B.P.

The basal sediments (below 44cm) in the 65cm-long core are grey black lacustrine clay (Shan et al., 1996), suggesting a moderate water depth. The presence of aquatic pollen *Typha* (60-55cm and 50-40cm) is consistent with the moderate water depth. A sample from a depth of 62cm has been radiocarbon-dated to 11,195±344 yr B.P. This unit is probably the same as the black lacustrine mud at the base of 60cm core, and thus confirms the existence of lacustrine conditions prior to 11,000 yr B.P. This suggests these basal units were formed at the same time as the +50cm high beach ridge.

The overlying sediment (44-22cm) is banded clay with light grey and light brown interbeds, suggesting an increase in water depth. The disappearance of aquatics may be consistent with deeper water. The unit is dated to between ca 8000 and 4000 yr B.P. by interpolation of the sedimentation rate. Shan et al. (1996) suggest that pollen assemblage from this interval is characteristic of mid-Holocene pollen assemblages from the region, and support a mid-Holocene attribution for this interval of increased water depth.

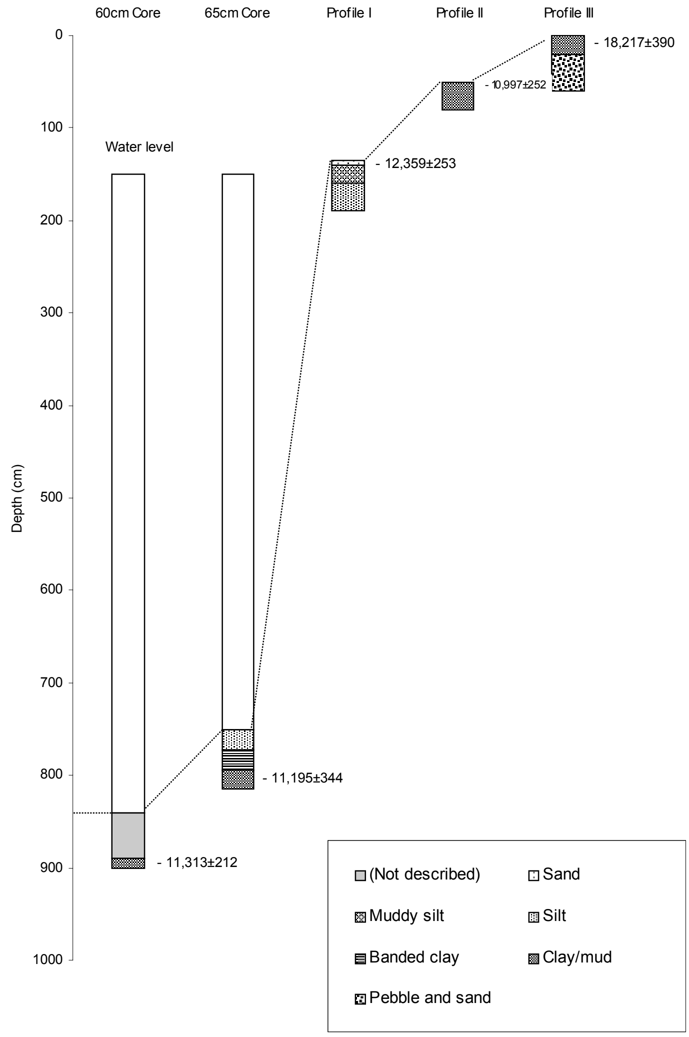
The uppermost sediment (above 22cm) is lacustrine silt, suggesting shallower conditions after ca 4000 yr B.P. The presence of abundant *Pediastrum boryanum* in the top of the unit (above 10cm) suggests a further decrease in water depth after ca 1800 yr B.P.

Hu (1995) has used the data from Wulanwula Lake, in conjunction with evidence from other nearby lakes, to reconstruct water-balance changes in the Kekexili Region since the last glacial maximum. The glacial maximum shoreline at Wulanwulu Lake can be correlated with similar features in nearby basins (less than 100km in distance): a lacustrine silt and clay unit (42-57cm above lake level) overlain by aeolian silt and sand from Goujumai Lake's terrace profile, radiocarbon-dated to 18,530±415 yr B.P.; and a lacustrine mud (ca 55cm above lake level) overlain by salt crust from Xiwujinlan Lake' terrace profile, U-series-dated to 22,100±1840 yr B.P. (Hu, 1995). Hu (1995) has suggested that there was a phase of high lake levels between 22,000-18,000 yr B.P. throughout the Kekexili Region.

The high lake level around 12,000 yr B.P. at Wulanwula Lake can be correlated to two nearby basins (less than 100km in distance): A nearshore sediment core from a 80cm-deep water in Jieyue Lake, shows a lacustrine blue-grey and yellow-brow banded mud with fine pebbles (0-24cm) and a lacustrine blue-grey clay (24-30cm). A sample at 24-30cm is radiocarbon-dated to 13,409±569 yr B.P. (Hu, 1995). The other sediment profile at 100cm above lake level with a yellow-brown lacustrine muddy silt (0.1-0.3cm) and lacustrine black silt (0.3-1.0cm), and grey-brown medium-fine sand (below 1.0cm) from Haiding Nur flat. A sample at 0.3-1.0cm is radiocarbon-dated to 13,618±299 yr B.P. Hu (1995) interpreted these units as marking a phase of high lake level during the late glacial in the Kekexili Region.

The high lake stand dated to ca 11,000 yr B.P. at Wulanwula Lake can be correlated with evidence from Zhuonai Lake and Changcuo (Hu, 1995). A sediment profile at 50cm above modern lake level in Zhuonai Lake Basin shows a brown lacustrine muddy silt and mud unit with abundant leaves of *Potamogeton* (0-48cm), a black lacustrine silt with remains of *Potamogeton* (48-63cm), and stone bedrock (below 63cm). A sample at 48-63cm is radiocarbon-dated to 10,124±228 yr B.P. The other sediment profile from the center of Changcuo, a 10cm-deep salt-lake shows a lacustrine mud (5-10cm) overlain by red-brown silt in 2cm thick and gypsum in 3cm thick. A sample from the lacustrine unit at 5-10cm is radiocarbon-dated to 9810±210 yr B.P. (Hu, 1995). Hu (1995) interpreted these units as marking a high lake level during the pre- and early Holocene in the Kekexili Region.

In the status coding, low (1) is indicated by lacustrine silt with abundant *Pediastrum boryanum* in near lake center core, and presumed to be related to modern lake level; moderately low (2) by silt deposit without *Pediastrum boryanum* in the central core, or grey muddy silt in Profile I, corresponding to the shorelines +15cm above lake level; intermediate (3) by non-laminated black mud or lacustrine clay in the central cores, corresponding to shorelines +50cm above lake level; moderately high (4) by laminated lacustrine mud or clay in the central cores; high (5) by 150cm shorelines above lake level in the basin.



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Radiocarbon dates

10,997±252	ca 20cm deep, lacustrine beach rock, Profile II.
11,313±212	50-60cm deep, mud, 60cm-long core
11,195±344	62cm deep, plant remains, 65cm-long core
12,359±253	5-10cm deep, muddy silt, Profile I
18,217±390	ca 10cm, lacustrine beach rock, Profile III.

(Lab numbers are not given; the samples were dated in Salt Lake Institute of Chinese Academy of Science)

Coding

18,500 -18,000 yr B.P.	high (5)
? -12,300 yr B.P.	moderately low (2)
11,400-8000 yr B.P.	intermediate (3)
8000-4000 yr B.P.	moderately high (4)
4000-1800 yr B.P.	moderately low (2)
1800-0 yr B.P.	low (1)

Preliminary coding: 03-11-1998
Final coding: 15-12-1998
Coded by GY and SPH

3.16. Salawusu Palaeolake, Shaanxi Province

Salawusu Palaeolake (also called Sjara Osso by Teilhard de Chardin and Licent, 1924; Chinese Geology Society, 1956; and Sala Us by Sun et al., 1996) (37.7°N, 108.6° E, 1300m above sea level) is a large former lake in the southern Eerduosi Plateau (the southern part of the Mongolian Plateau). The palaeolake area, which had an area of ca 100 km^2, was located in the boundary zone between the Loess Plateau (to the south of Salawusu) and the Maowusu Desert on the Eerduosi Plateau (to the north of Salawusu) and was therefore very sensitive to climatic changes (Dong et al., 1983; Zheng, 1989). The palaeolake basin originated as a tectonic depression and the bedrock is Cretaceous red sandstone (Dong et al., 1983). The basin today is drained by the Salawusu River, which is a tributary of the Yellow River. The Yellow River flows in a major arc around the Eerduosi Plateau, and is joined by the Salawusu River as it flows southwards again towards the Loess Plateau. The Salawusu River has cut down through the lacustrine deposits of the Salawusu Palaeolake, forming 60-70m high terraces (Dong et al., 1983; Zheng, 1989). Neotectonism within the last 2000 years appears to be responsible for the shift from lacustrine and/or fluvial deposition to fluvial erosion. However, it is generally assumed that there was little or no active tectonism during the Late Pleistocene. The major sediments within the palaeolake basin today are aeolian sand dunes on the top of the river terraces, and fluvial sands in the Salawusu River valley (Dong et al., 1983). The regional climate is semi-arid, with annual temperatures of 4-14°C and annual precipitation of 250-450mm (Gao et al., 1985).

Teilhard de Chardin and Licent (1924) investigated the Upper Pleistocene Salawusu fluvial-lacustrine strata in which abundant animal remians and stone artifacts were found. There have been many subsequent studies on these Upper Pleistocene fluvial-lacustrine strata, and these units have become a type section for the Upper Pleistocene in North China (Chinese Geology Society, 1956). There have also been many studies on the Late Quaternary history of the basin (e.g. Pei and Li, 1964; Qi, 1975; Yuan, 1978; Dong et al., 1983; Yuan, 1988; Zheng, 1989; Su and Dong, 1994). During the early part of the Late Pleistocene, the Salawusu was characterised by an alternation of fluvial and aeolian deposition. Subsequently the basin became closed and a lake formed. There was a stable lacustrine phase during the middle part of the Late Pleistocene. The later part of the Late Pleistocene was characterised by aeolian deposition. Lacustrine deposition occurred during the early and mid-Holocene, and alternation of fluvial and aeolian deposition is characteristic of the late Holocene. Thus, the Sulawusu Palaeolake existed during the middle Lower Pleistocene and the early to mid Holocene, when the climate was very wet and the deserts retreated northwards from the Salawusu area (Yuan, 1988; Zheng, 1989; Sun et al., 1996). The Salawusu River was re-established ca 2300 yr B.P. and has drained the basin since then.

A classic profile through a 60m-high terrace in the Salawusu River bank, at Dishaogouwan (Dishaogouwan Profile), provides a sedimentary record back to the early Late Pleistocene. The earlier work on this profile (before 1980's) mostly focussed on the stratigraphy and faunal biostratigraphy. Recent work has examined the mineralogy (e.g. Lu, 1985) and geochemistry (e.g. Gao et al., 1985) of the deposits, and established a chronology based of ^{14}C (Li et al., 1984; Su and Dong, 1994) and TL (Zheng, 1989; Li et al., 1993; Sun et al., 1996; Su and Dong, 1997) dating. Changes in lake status are reconstructed on the basis of changes in lithology, clay mineral assemblages, geochemistry, and the presence or absence of mollusc shells, in the Dishaogouwan

Profile. The interpretation of the changes in lake status generally follows the reconstructions put forward by Zheng (1989) and Sun et al. (1996). The chronology is based on 10 radiocarbon-dates (Li, et al., 1984; Su and Dong, 1994) and 6 TL dates (Zheng, 1989; Li et al., 1993; Sun et al., 1996; Su and Dong, 1997) from the profile.

The Dishaogouwan Profile was originally described by Zheng (1989). A more detailed description of part of the profile was given by Sun et al. (1996). The composite profile is described here. The Dishaogouwan Profile (below 59.3m) bottoms out on Cretaceous red sandstone. The basal sediments (59.3-44.3m) consist of an alternation of yellow fine sand with crossbedding, characteristic of aeolian deposition (59.3-52.5m, 51.5-49.2m, 47.0-46.0m), and pale-white bedded silt, characteristic of fluvial deposition (52.5-51.5m, 49.2-47.0m, 46.0-44.3m). The famous Salawusu mammal fossils (Tailard de Chardin and Licent, 1924) were chiefly found in the fluvial deposits between 49.2-47.0m. Samples from ca 58.0-58.5m and ca 44.5-45.0m were TL-dated to 216,000±22,000 and 177,000±14,000 yr B.P. respectively, suggesting that this fluvial-aeolian phase occurred between ca 220,000-170,000 yr B.P.

The overlying sediment (44.3-22.4m) is lacustrine clay, indicating that the lake basin became closed. There is no geomorphic evidence indicating why the basin became closed. However, the fact that lacustrine conditions persisted for a long period indicates that climatic conditions were consistent with the existence of a lake and therefore must have been considerably wetter than today. The dominance of kaolinite in the clay mineral assemblage is consistent with humid conditions (Lu, 1985), as is the relatively low ratio of SiO_2/Al_2O_3 (< 12) (Gao et al., 1985). The unit can be sub-divided, on the basis of changes in lithology, into three sub-units. The lower part of the unit (44.4-26.8m) is pale-green to pale-yellow laminated silt and clayey silt. The existence of laminations indicates deepwater conditions. Samples from 44.0-44.5m, 43.0-43.5m and 37.5-38.0m were TL-dated to 136,000±15,200, 124,900±15,200 and 93,000±14,000 yr B.P., suggesting this deepwater lake occurred between ca 170,000-89,000 yr B.P. The middle part of the lacustrine unit (26.8-25.0m) is a layer of laminated silt containing freshwater mollusc shells. Although the preservation of laminations indicates the lake remained relatively deep, the occurrence of mollusc shells suggests it was shallower than before. This unit is dated to between ca 89,000-80,000 yr B.P. by interpolation of sedimentation rate (0.021 cm/yr) between the TL-dates from 37.5-38.0m and 23.0m. The upper part of the unit (25.0-22.4m) is sand and silt, with gently-inclined bedding. The bedding is is characteristic of nearshore deposition, suggesting water depth decreased further. A sample from ca 23.0m was TL-dated to 70,900±6200 yr B.P., suggesting this nearshore unit was deposited between ca 80,000-70,000 yr B.P.

The overlying sediment (22.4-4.0m) is well-sorted fine sand with steep-angled (20-31°) diabedding and crossbedding, characteristic of aeolian deposition.The clay mineral assemblage is characterised by illite, and the absence of kaolinite and montmorillinite, which is consistent with aeolain deposition (Lu, 1985). The relatively high ratio of SiO_2/Al_2O_3 (17-20) is consistent with dry conditions (Gao et al., 1985). The presence of aeolian sands indicates that the lake was considerably smaller than before, and may even have dried out. Two layers of lacustrine sandy silt have been identified within this primarily aeolian unit (Sun et al., 1996). Both are characterised by kaolinite and low ratios (<15) of SiO_2/Al_2O_3. The lower layer of lacustrine sediment (13.9-13.5m) is pale-green sandy silt with abundant freshwater mollusc shells, suggesting relatively shallow conditions. A sample from 13.60-13.65m was radiocarbon-dated to

30,240±1280 yr B.P., suggesting this shallow water phase occurred between 32,000-30,000 yr B.P. The upper layer of lacustrine sediment (10.5-12.0m) is also pale-green sandy silt, but does not contain freshwater mollusc shells. This suggests water depth was greater than during the first lacustrine phase. Samples from ca 10.5m and ca 11.5m were radiocarbon-dated to 27,940±600 and 28,170±1080 yr B.P. respectively, indicating that the second lacustrine phase occurred between ca 27,900-28,300 yr B.P.

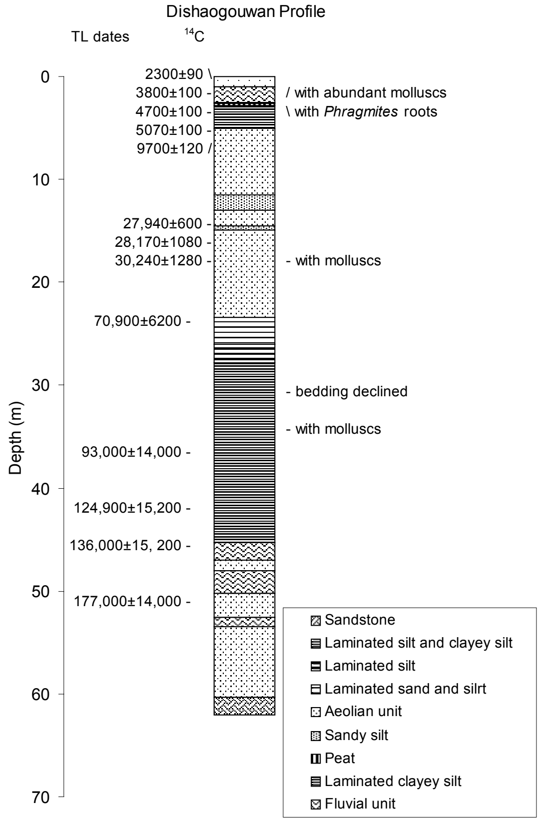
The overlying sediment (4.0-1.80m) is grey-blue laminated clayey silt of lacustrine origin, indicating a return to deep water conditions after the aeolian phase. The unit is characterised by kaolinite and low ratios of SiO_2/Al_2O_3 (< 10), consistent with lacustrine deposition. The bottom boundary of this unit is dated by three radiocarbon samples (9700±120, 9600±100, 9500±100) and the top boundary has been dated to 5070±100 yr B.P. This deepwater lacustrine phase is therefore generally assumed to have occurred between ca 10,000-5000 yr B.P. (Zheng, 1989; Sun et al., 1996).

The overlying sediment (1.65-1.80m) is peat with abundant *Phragmites* roots. This unit is characteristic of swamp deposition, and indicates shallowing of the lake after ca 5000 yr B.P.

The overlying sediment (1.65-1.50m) is grey-blue laminated clayey silt with abundant mollusc shells, indicating a return to lacustrine conditions. Samples from the bottom and the top of this unit were radiocarbon-dated to 4700±100 and 3800±100 yr B.P. respectively, indicating that the return to lacustrine conditions occurred between 4700-3800 yr B.P.

The uppermost sediment (1.5-0m) is light-yellow sandy silt with characteristic fluvial bedding. The presence of fluvial sediments indicates that the lake became open and started to overflow via the Salawusu River. The disappearance of kaolinite from the clay mineral assemblage and the increased ratios of $SiO_2/A_{l2}O_3$ (ca 17) are consistent with drier conditions. A sample from this top unit was radiocarbon-dated to 2300±90 yr B.P. Deposition ceased after 2300 yr B.P., and the Salawusu River began to cut down through the former lake sediments. This shift to fluvial erosion is thought to be related to renewed uplift of the Eerduosi Plateau (Dong et al., 1983; Sun et al., 1996) after ca 2300 yr B.P.

In the status coding, very low (1) is indicated by aeolian sand deposition or modern conditions; low (2) by peat deposition; moderately low (3) by nearshore lacustrine sand and silt deposits, with gently-inclined bedding; intermediate (4) by lacustrine sandy silt with abundant mollusc shells; moderately high (5) by lacustrine sandy silt without molluscs shells; high (6) by laminated silty clays or clays with molluscs; and very high (7) by laminated silty clays or clays without molluscs.



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Radiocarbon dates

30,240±1280	ca 13.60-13.65m, Dishaogouwan Profile
28,170±1080	ca 11.5m, Dishaogouwan Profile
27,940±600	ca 10.5m, Dishaogouwan Profile
9700±120	ca 4.0-3.5m, Dishaogouwan Profile
9600±100	ca 4.0-3.5m, Dishaogouwan Profile
9500±100	ca 4.0-3.5m, Dishaogouwan Profile
5070±100	ca 1.8m, Dishaogouwan Profile
4700±100	ca 1.6m, Dishaogouwan Profile
3800±100	ca 1.5m, Dishaogouwan Profile
2300±90	ca 0.5m, Dishaogouwan Profile

TL-dates

216,000±22,000	ca 58.0-58.5m, Dishaogouwan Profile
177,000±14,000	ca 44.5-45.0m, Dishaogouwan Profile
136,000±15,200	ca 44.0-44.5m, Dishaogouwan Profile
124,900±15,200	ca 43.0-43.5m, Dishaogouwan Profile
93,000±14,000	ca 37.5-38.0m, Dishaogouwan Profile
70,900±6200	ca 23.0m, Dishaogouwan Profile

Coding

220,000-170,000 yr B.P.	not coded (prior to formation of palaeolake)
170,000-89,000 yr B.P.	very high (7)
89,000-80,000 yr B.P.	high (6)
80,000-70,000 yr B.P.	moderately low (3)
70,000-32,000 yr B.P.	very low (1)
32,000-30,000 yr B.P.	intermediate (4)
30,000-28,300 yr B.P.	very low (1)
28,300-27,900 yr B.P.	moderately high (5)
27,900-9700 yr B.P.	very low (1)
9700-5000 yr B.P.	very high (7)
5000-4700 yr B.P.	low (2)
4700-3800 yr B.P.	high (6)
3800-100 yr B.P.	not coded (fluvial deposition and fluvial erosion)
0 yr B.P.	very low (1)

Preliminary coding: 11-02-1999
Final coding: 27-03-99
Coded by GY and SPH

3.17. Shayema Lake, Sichuan Province

Shayema Lake (28°50'N, 102°12'E, 2400 m a.s.l.) lies in the southwestern part of Sichuan province. In September 1987, the lake had an area of 0.04 km^2, and an average water depth of 10 m (Jarvis, 1993). The lake is thought to have been formed when a former river course was blocked by faulting (Jarvis, 1993). The region is still prone to earthquakes. The lake is currently fed by some small streams draining the catchment and by direct precipitation. The annual precipitation is ca 800-1200 mm, 80% of which occurs during the summer as a result of the penetration of the southeastern and southwestern monsoons into the region. The Tibetan and Mongolian high pressure systems prevent the penetration of rain-bearing winds into this region during the winter half year.

Two cores, 11 m (Core 1) and 9 m (Core 2) long respectively, were taken close together from the centre of the lake. The top metre of the cores was not analysed because of mixing of the unconsolidated sediments during transport (Jarvis, 1993). Although both cores appear to have been sampled, the stratigraphic and palynological data given in Jarvis (1993) are only for the 11m long core (Core 1). We assume that both cores show a similar record. Core 1 provides a record back to ca 10800 yr B.P. (Jarvis, 1993). Changes in relative water depth are based on changes in lithology and loss-on-ignition, and broadly follow the interpretation of the original author. The chronology is based on five radiocarbon dates, all of which appear to have come from Core 1.

The basal unit (11.0-9.2 m) is fine detrital gyttja and fine clay. The loss-on-ignition is relatively low (ca 20%) throughout the unit. The lithology and the loss-on-ignition are consistent with moderately deep water conditions. A sample from ca 10.5 m is radiocarbon dated to 10770±90 yr B.P. By linear interpolation of the sedimentary rate (1.17 mm/yr) between this date and a date from ca 7.85-7.8 m, this phase of moderately deep water conditions occurred between ca 10800 and 8960 yr B.P., although the original author has suggested this phase occurred 10800-9100 yr B.P.

The overlying unit (9.2-7.85 m) is fine detrital gyttja. The loss-on-ignition increases to 30-50%. The change in lithology and the increase in organic content are both consistent with a decrease in water depth. By linear interpolation, this interval of shallower conditions occurred ca 8960-7800 yr B.P.

The overlying unit (7.85-7.8 m) is yellowish clay. The change in lithology could be consistent with an increase in water depth. However, the loss-on-ignition is high (60%), which is inconsistent with increased water depth. Given the thinness of this unit, the colour of the clay which is consistent with subaerial derivation, and the fact that the overlying unit reflects very shallow conditions, we interpret this yellowish clay unit as being formed by subaerial erosion into the lake probably associated with the catastrophic events associated with the subsequent lake lowering.

The overlying unit (7.8-6.75 m) is mossy peat, indicating a shift to shallow water conditions. The loss-on-ignition values for this unit are ca 65%, consistent with shallow water conditions. The transition between this unit and the underlying lacustrine units is abrupt. Javis (1993) suggests that there was a sudden change from deep to shallow conditions, ca 7700 yr B.P. Given that the terrestrial pollen record does not show a similarly abrupt change across this boundary, she suggests that this change in water depth was the result of tectonic movements (earthquakes). The abrupt transition from gyttja, via clay to mossy peat is consistent with a catastrophic draining of the lake due to tectonic movements. However, the persistence of mossy peat deposition for a period of ca 1500 years (ca 7700-6100 yr B.P.) indicates that the lake level remained at its new lower level for a considerable time, presumably as a result of a significant reduction in the regional water budget. The terrestrial pollen record shows a gradual transition between more mesic to more sclerophyll tree species between ca 7700-5300 yr B.P., consistent with the idea that although the initial change in lake depth might have been due to a catastrophic event, the persistence of shallow conditions was the result of climate change.

The overlying unit (6.75-4.43 m) is mossy peat containing increased amounts of fine detrital gyttja. The increased gyttja content indicates an increase in water depth. The loss-on-ignition decreases to ca 50%, consistent with increased water depth. Two samples from within the unit, at ca 6.2 m and 4.45 m, are radiocarbon dated to 5290±100 and 2840±60 yr B.P. respectively. By interpolation of the sedimentary rate (0.714 mm/yr) between these two dates, this interval of increased water depth occurred ca 6100-2800 yr B.P. There are three thin layers of yellow clay, fine sand and coarse sand within the top part of the unit at 5.40 m, 5.20 m and 4.60 m respectively. Each of these units is marked by an increase in loss-on-ignition. Jarvis (1993) interprets these units as marking episodes of erosion from the catchment area, rather than fluctuations in water depth.

The overlying unit (4.43-3.6 m) is fine-grained detrital gyttja, marking a further increase in water depth. However, the loss-on-ignition increased to between 60-70%, which is apparently inconsistent with our interpretation that loss-on-ignition decreases as water depth increases. However, the increase in loss-on-ignition coincides with an expansion of riparian vegetation and may simply reflect local changes in vegetation cover. By interpolation of the sedimentary rate (1.16 mm/yr) between the radiocarbon dates from 4.45 m and 2.35 m, this unit was formed between ca 2800 and 2110 yr B.P.

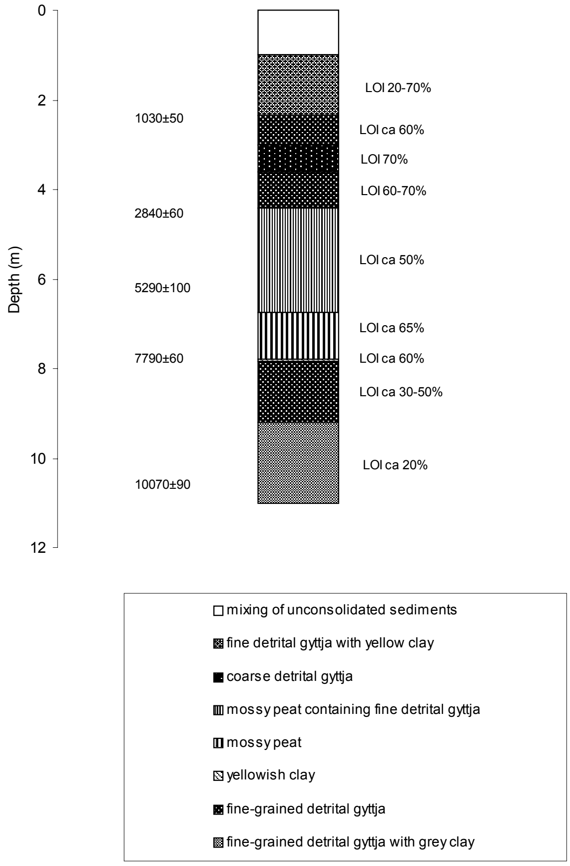
The overlying unit (3.6-3.0 m) is coarse detrital gyttja. The increase in coarse matter suggests that the water depth decreased. The loss-on-ignition are high (70%), consistent with a decrease in water depth. This phase occurred between ca 2110 and 1590 yr B.P.

The overlying unit (3.0-2.34 m) is fine-grained detrital gyttja, indicating that water depth increased. A decrease in loss-on-ignition to ca 60% is consistent with increased water depth. A sample from ca 2.35 m is radiocarbon dated to 1030±50 yr B.P. Linear interpolation between this date and the date from 4.45 m suggests that this phase of increased water depth occurred ca 1590-1000 yr B.P.

The overlying unit (2.34-1.0 m) consists of alternating bands of fine detrital gyttja and yellowish clay. Jarvis (1993) interpreted the abundance of clay laminae as a reflection of increased erosion into the lake as a consequence of human disturbance of the catchment. The values for loss-on-ignition show marked fluctuations within this unit, with the highest values coincident with the yellow clay layers. Marked fluctuations in loss-on-ignition are consistent with high inputs of eroded material, and probably do not reflect any significant change in water depth compared to the period before ca 1000 yr B.P.

In the status coding, very low (1) is indicated by moss peat deposition after the catastrophic lowering of the lake, low (2) by moss peat with fine detrital gyttja; intermediate (2) by coarse detrital gyttja; high (4) by fine detrital gyttja. The interval before ca 7700 yr B.P. is not coded because it is impossible to estimate the water depth equivalence of the units before and after the catastrophic lowering. The period after

1000 yr B.P. is coded as a continuation of the lake status between 1590-1000 yr B.P. on the assumption that the change in lithology reflects increased anthropogenic disturbance of the catchment.



Reference:

Jarvis DI (1993) Pollen evidence of changing Holocene monsoon climate in Sichuan Province, China. Quaternary Research 39: 325-337

Radiocarbon dates

10,070±90	ca 10.5 m
7790±60	ca 7.85-7.8 m
5290±100	ca 6.2 m
2840±60	ca 4.45 m
1030±50	ca 2.35 m

Coding

Before 7700 yr B.P.	not coded due to the uncertainty of the water-depth equivalence
7700-6100 yr B.P.	very low (1)
6100-2800 yr B.P.	low (2)
2800-2110 yr B.P.	high (4)
2110-1590 yr B.P.	intermediate (3)
1590-0 yr B.P.	high (4)

Preliminary coding: January 1999
Final coding: February 1999
Coded by SPH and BX

3.18. Big Ghost Lake, Taiwan

Big Ghost Lake (22° 51' 15" N, 120° 51‘ 15" E, 2150 m a.s.l.) is a relatively small subalpine lake in the Central Range, central Taiwan, immune from anthropogenic influence (Luo et al., 1996). The lake area is 0.1087 km^2 (Luo, 1996). The lake has a maximum depth of ca 40m, a mean depth of 15.4m, and the deepwater below the thermocline is anoxic throughout most of the year. The lake is fed by direct precipitation and surface runoff from the limited catchment (ca 0.9km^2), and has no outflow. The lake level fluctuates ca 2m between the dry and rainy seasons. The basin is of tectonic origin, but there is no evidence of recent tectonism. The bedrock in the basin is Miocene argillite and slate.The annual mean temperature is ca 13° C and the annual mean precipitation is ca 4200 mm. The vegetation close to the lake basin is moss-covered alpine forest.

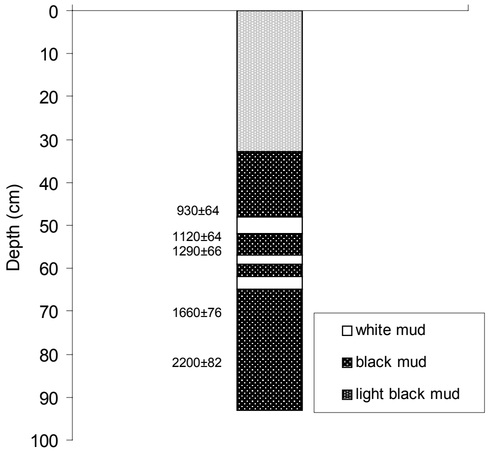
A 93-cm long core, taken from the centre of the lake in a water depth of ca 30m, provides a lithological record back to ca 2600 yr B.P. (Luo et al., 1996). The lake sediments are not annually-laminated, but distinct horizontal bedding is clearly preserved, indicating that the lake has been relatively deep throughout the last 2600 years. The sediments are of two distinct types: dark black or brown units alternate with pale-coloured or white sediments. The boundary between the units is always sharp and distinct. The dark-coloured sediment has a mean grain size of ca 52µm, is relatively rich in carbon (mean value ca 16.3%, range 10-20%) and sulphur (0.15%). The light-coloured units are finer-grained (mean grain size 9.9µm), contain relatively little carbon (ca 3.5%, range 2-3%) and sulphur (ca 0.026%). Twigs and small fragments of leaves frequently occur at the lower boundary of the pale-coloured units. Luo et al. (1996) have suggested that the dark-coloured units were formed under anoxic conditions in relatively deep water (i.e. water depths similar to today), and that the pale-coloured units mark intervals when the climate was colder and drier, and lake level was lower than today. Luo and Chen (1998), on the basis of an evaluation of the elemental composition of the sediments, have suggested that the water depth must have been at least 2m less when the pale-coloured units were formed than when the dark-coloured units were formed. Given the great depth of this lake, it seems that significantly larger changes in water depth would be required to produce oxic conditions in the deepest part of the lake, and it seems likely that the transition from dark- to light-coloured material must indicate more pronounced lake-level changes than Luo and Chen (1998) would imply.

The chronology of water depth changes at Big Ghost Lake is based on AMS-radiocarbon and Pb^{210} dating. The Pb^{210} dating indicates that the sedimentation rate in the uppermost part of the core is 0.48 mm/yr, and extrapolation of this rate downcore yields a date consistent with that obtained for the uppermost radiocarbon sample (47-48 cm). There are five radiocarbon dates on the lower part of the profile. The sedimentation rate on the lower part of the profile appears to have been relatively constant (between 0.024 and 0.036cm/yr). However, the chronology used here is based on linear interpolation between the available dates rather than the use of an average sedimentation rate.

Dark-coloured units occur between 93-84cm, 83-79 cm, 78-75cm, 74-65cm, 62-59cm, 57-52cm and 48-0cm. These units indicate deepwater phases between 2615-2240 yr B.P., 2200-2035 yr B.P., 1990-1870 yr B.P., 1825-1525 yr B.P., 1445-

1360 yr B.P., 1305-1100 yr B.P., and 950-0 yr B.P. The intervening units are pale-coloured muds, and indicate shallower-water phases.

In the status coding, low (1) is indicated by pale-coloured deposits, and high (2) by dark-coloured deposits.



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Radiocarbon Dates

Non-calibrated	Calibrated	
	2200±82	83 cm, organic component (AMS)
	1660±76	70 cm, organic component (AMS)
	1290±66	57-56 cm, organic component (AMS)
	1120±64	52-53 cm, organic component (AMS)
	930±64	47-48 cm, organic component (AMS)

Coding

2615-2240 yr B.P.	high (2)
2240-2200 yr B.P.	low (1)
2200-2035 yr B.P.	high (2)
2035-1990 yr B.P.	low (1)
1990-1870 yr B.P.	high (2)
1870-1825 yr B.P.	low (1)
1825-1525 yr B.P.	high (2)
1525-1445 yr B.P.	low (1)
1445-1360 yr B.P.	high (2)
1360-1305 yr B.P.	low (1)
1305-1100 yr B.P.	high (2)
1100-950 yr B.P.	low (1)
950-0 yr B.P.	high (2)

Preliminary coding: January 1999
Final coding: March 1999
Coded by BX and SPH

3.19. Chitsai Lake, Taiwan

Chitsai basin (23°45'10''N, 121°14'10''E, 2890 m a.s.l.) lies in the Central Range of central Taiwan. It is an alpine lake immune from anthropogenic influence (Liew and Huang, 1994). There are two lakes in the Chitsai basin: a big one covering an area of 22000 m^2, and a small one which has an area of 6000 m^2. The two lakes are 60 m apart. The smaller lake has become smaller and shallower in recent years, but has not become completely desiccated (Liew, pers. comm., 1999). The lakes are fed by direct precipitation. The bedrock in the basin is schist of the metamorphic belt of the Central Range. *Tsuga-Picea* forests surrounded the lake.

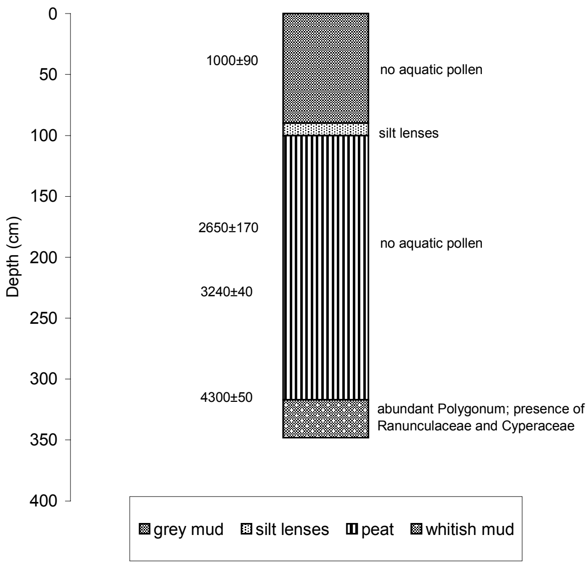
A 348 cm core taken from a recently desiccated part of the smaller lake provides a sedimentary record back to ca 5300 calendar yr B.P., 4700 ^{14}C yr B.P. (Liew and Huang, 1994; Liew, pers. comm., 1999). Changes in water-depth are based on changes in lithology and the presence/absence of aquatic pollen, and follow the interpretation of the original authors. Four calibrated radiocarbon dates are given in the published paper (Liew and Huang, 1994); the chronology used here is based on the non-calibrated ages for these four radiocarbon dates (Liew, pers. comm., 1999).

The basal unit (348-317 cm) is whitish mud. The fine-grained lithology, with an average diameter ca 10-20 µm, suggests deep water conditions. The abundance of *Polygonum* (ca 5-10%), and the presence of Ranunculaceae and sparse Cyperaceae are consistent with relatively deep water. A sample from the base of the overlying unit (317-312 cm) was radiocarbon dated to 4300±50 yr B.P. Another sample from 220-230 cm was dated to 3240±40 yr B.P. By extrapolation of the sedimentation rate from the overlying sediments, this phase of relatively deep water conditions occurred ca 4700-4340 yr B.P.

A shallowing of the lake is indicated by peat deposition between 317-100 cm. The non-organic component of the sediment is coarser than the underlying unit, with an average diameter of 30-40 µm, consistent with shallowing. The absence of aquatic pollen is consistent with shallowing. Two samples from 220-230 cm and 170 cm were radiocarbon dated to 3240±40 and 2650±60 yr B.P. respectively. Using the sedimentation rate of 0.93 mm/yr between these dated samples, this phase occurred ca 4340-1900 yr B.P.

The overlying unit (100-0 cm) is grey mud, suggesting a return to deeper water. The average diameter of the non-organic component of the sediment is finer (ca 15-40 µm), consistent with deepening. However, there is no aquatic pollen, suggesting that the water depth was not as great as during the deposition of the white mud. There are discrete lenses of silt incorporated in the basal 10 cm of this unit (Liew and Huang, 1994). These silt lenses may have resulted from a storm event (Liew, pers. comm., 1999), or through erosion and incorporation of subaerial material as the water level increased. A sample of wood from 35 cm was radiocarbon dated to 1000±90 yr B.P. This phase of lacustrine conditions occurred ca 1900-0 yr B.P.

In the status coding, low (1) is indicated by peat; intermediate (2) by grey mud; and high (3) by whitish mud.



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Liew, P.M., 1999. Personal communication (letter)

Radiocarbon dates

		Calibrated ages	
NTU-1746	4300±50	4840±50	312-317 cm, peat
NTU-1733	3240±40	3467±40	220-230 cm, peat
NTU-1729	2650±60	2763±60	170 cm, peat
NTU-1732	1000±90	946±90	35 cm, wood

Coding

4700-4340 yr B.P.	high (3)
4340-1900 yr B.P.	low (1)
1900-0 yr B.P.	intermediate (2)

Preliminary coding: December 1998
Final coding: February 1999
Coded by BX and SPH

3.20. Toushe Lake, Taiwan

Toushe Lake (23.82°N, 120.89°E, 650m above sea level) is a former lake basin in the hilly area of central Taiwan. The lake dried up ca 1500 yr B.P. (Liew et al., 1998). The area of the former lake, as indicated by the extent of peat and gyttja deposits within the basin, was ca 0.1km^2. The basin originated as a fault depression during the late Pleistocene. The bedrock is Tertiary slate (Liew et al., 1998). The basin lies within the region influenced by the Pacific summer monsoon and has a charactaristically warm and wet climate, with mean annual temperature of 19.2°C, total annual precipitation of 2341mm and annual evaporation of 1098mm (Liew et al., 1998). The vegetation in Toushe basin is evergreen forest, dominated by *Machilus* and *Castanopsis* (Huang et al., 1997).

The uppermost 17m of a ca 40m-long core, taken from the centre of the basin, provides a sedimentary record back to 30,000 yr B.P. (Huang et al., 1997; Liew et al., 1998). There is one AMS-radiocarbon date and 15 conventional radiocarbon dates from the uppermost 16.90m of the core. The core material is predominantly peat, although there are some gyttja units, and thin (2-4cm thick) clay bands occur within the peat. Huang et al. (1997) and Liew et al. (1998) have reconstructed changes in palaeoenvironmental and palaeoclimatic conditions during this interval on the basis of changes in the terrestrial pollen assemblages, sedimentation rates, organic content as measured by loss-on-ignition, and the frequency of clay layers in the core sequence. The clay layers are supposed to have formed as a result of the deposition of clastics, derived by erosion from the catchment, during heavy rain events during the summer monsoon season (Huang et al., 1997). The frequency of clay layers within the deposits are therefore interpreted as an indicator of the frequency of extreme rainfall events. On the basis of the pollen and lithological data, Huang et al. (1997) and Liew et al. (1998) suggested that conditions in the last glacial (ca 25,000-12,000 yr B.P.) were generally colder and drier because the winter monsoon was stronger than today. They suggest there was an abrupt change to wetter and warmer conditions during the late glacial (ca 12,000-10,000 yr B.P.), and peaking in the Holocene because of the strengthening of the summer monsoon.

The presence of discrete gyttja units within the core material imply that there was open water within the Touche basin at certain intervals, even though the sedimentary records indicate that the basin was occupied by a peat bog most of the time. Neither Huang et al. (1997) nor Liew et al. (1998) comment on the environmental significance of the gyttja layers. Here, we use the lithological changes from gyttja to peat, supported by the loss-on-ignition record and sedimentation rates, to reconstruct changes in water depth in the former Toushe Lake over the last 30,000 years. We assume that the clay layers represent short-lived erosion events, as postulated by Liew et al. (1998), and therefore provide no indication of changes in water depth.

The basal sediments (ca 17.2-14.1m) are peat, containing two distinct clay layers. The presence of peat indicates relatively shallow water conditions. The high loss-on-ignition values (60-90%) are consistent with shallow water. Two samples from 16.80-16.90m and 15.87-15.97m within the unit were radiocarbon-dated to 29,300±300 and 28,000±250 yr B.P. respectively. Extrapolation of the sedimentation rate (0.0715 cm/yr) between these two radiocarbon dates, suggest that the unit was formed between ca 30,000 and 25,460 yr B.P. Extrapolation of the sedimentation rate (0.0415 cm/yr) between the uppermost of these radiocarbon dates (15.87-15.97m) and a date of 18,130±160 yr B.P. from a depth of 11.77-11.87m would yield an age of 23,620 yr B.P. for the end of this phase of peat deposition. Given the fact that there are several different lithological units, each with potentially very different sedimentation rates, present between the basal peat and the next radiocarbon-dated unit, the date of 25,460 yr B.P. is the most likely estimate for the cessation of peat deposition.

The overlying sediments (ca 14.1-14.0m) are gyttja. This change in lithology indicates a short-lived increase in water depth. A decrease on organic content, as registered by loss-on-ignition (50%), is consistent with increasing water depth. The exact timing of this short-lived increase is uncertain. Estimates based on extrapolation of the sedimentation rate from the underlying peat would date the interval to 25,460 to 25,320 yr B.P., whereas estimates based on interpolation of the sedimentation rate between the radiocarbon samples loosely bracketing the unit would date it to 23,620 to 23,370 yr B.P.

The overlying unit (ca 14.0-13.6m) is peat. The change in lithology indicates a return to shallow water conditions. The increase in loss-on-ignition (90%) is consistent with this interpretation. This interval of relatively shallow conditions is dated by extrapolation of the sedimentation rate on the lowest peat to 25,320-24,760 yr B.P. and by interpolation between the loosely bracketing dates to 23,370-22,410 yr B.P.

The overlying sediments (ca 13.6-12.7m) are gyttja. This change in lithology indicates an increase in water depth. A decrease of organic content, as registered by loss-on-ignition (20-55%), is consistent with increasing water depth. This interval of relatively deep conditions is dated by extrapolation of the sedimentation rate on the lowest peat to 24,760-23,500 yr B.P. and by interpolation between the loosely bracketing dates to 22,410-20,240 yr B.P.

The overlying unit (ca 12.7-12.0m) is peat. The change in lithology indicates a return to shallow water conditions. The increase in loss-on-ignition (55-85%) is consistent with this interpretation. This interval of relatively shallow conditions is dated by extrapolation of the sedimentation rate on the lowest peat to 23,500-22,520 and by interpolation between the loosely bracketing dates to 20,240-18,550 yr B.P.

The overlying sediments (ca 12.0-11.6m) are gyttja. This change in lithology indicates an increase in water depth. A decrease in organic content, as registered by loss-on-ignition (20-55%), is consistent with increasing water depth. There is a radiocarbon date from this unit (11.77-11.87m) of 18,130±160 yr B.P. Using the sedimentation rate of 0.0415 cm/yr estimated between this date and the underlying date of 28,000 yr B.P. from the basal peat would suggest that the beginning of gyttja deposition occurred ca 18,550 yr B.P. Interpolation of the sedimentation (0.0426 cm/yr) between the date from the gyttja and a date of 12,350±90 yr B.P. from near the top of the overlying unit (9.3-9.41m) would indicate that gyttja deposition ceased ca 17,620 yr B.P. Thus, the lithological evidence suggests that there was an interval of ca 900-1000 yr centred of 18,000 yr B.P. that was wetter than during the remainder of the glacial interval.

The overlying unit (11.6-8.4m) is peat. There is a single clay unit present close to the top of the peat. The change in lithology indicates a return to shallower water conditions, as does the increase in organic content (loss-on-ignition of 65-95%). Two samples from near the top of the peat unit (9.3-9.41m and 8.61-8.70m) have been radiocarbon-dated to 12,350±90 and 12,100±90 yr B.P. respectively. Interpolation of the sedimentation rate

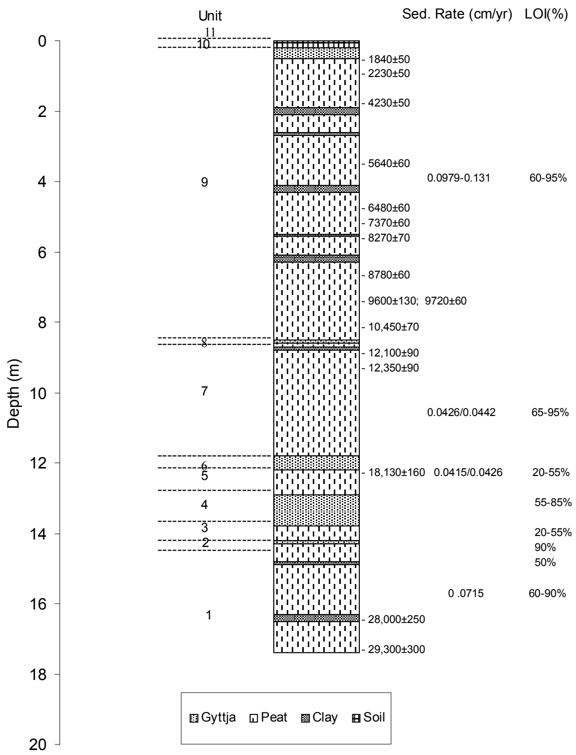
(0.0426 cm/yr) between the lowermost of these dates and the date from the underlying gyttja would suggest that the transition to peat deposition occurred 17,620 yr B.P. The sedimentation rate within the peat, as calculated from the two dates on the peat itself, is extremely high (0.22 cm/yr). Interpolation of this sedimentation rate would yield a date of 13,370 yr B.P. for the onset of peat deposition and 11,920 yr B.P. for its termination. Interpolation of the sedimentation rate (0.0442 cm/yr) between the uppermost radiocarbon date from the peat and a date of 10,450±70 yr B.P. from an overlying peat unit yields an estimate of 11,510 yr B.P. for the cessation of peat deposition. There is thus good agreement between the two estimates for the cessation of peat deposition, but a significant discrepancy between the two estimates for the onset of peat deposition. The estimate for the beginning of peat deposition based on the dates from the peat itself is unrealistically young. This suggests that there must have been a significant increase in sedimentation rate within the peat unit itself. Such an increase in sedimentation rate within the peat unit is consistent with a trend towards shallowing upcore.

The overlying sediment (ca 8.4-8.3m) is gyttja. The change in lithology indicates a short-lived interval of deeper water conditions, which started sometime between 11,920 and 11,510 yr B.P. and finished ca 11,290 yr B.P.

The overlying unit (8.3-0.3m) is peat, intercalated with at least five thin clay units. The change in lithology is consistent with a return to shallow-water conditions. Although the loss-on-ignition within the clay bands is low (<30%), the organic content of the peats is high (60-95%) consistent with shallow conditions. There are 11 radiocarbon dates from the peat. The uppermost sample, from a depth of 0.3-0.4m was radiocarbon-dated to 1840±50 yr B.P. The sedimentation rates between the radiocarbon dates range from 0.131 cm/yr to a minimum of 0.0979 cm/yr, consistent with relatively shallow conditions, but there is no systematic change in sedimentation rate that might indicate systematic changes in water depth. This interval of peat deposition can be dated to between 11,290 yr B.P. and 1800 yr B.P.

There are discrepancies in the description of the uppermost sediment (0.30-0m) given by Huang et al. (1997) and Liew et al. (1998). According to Huang et al. (1997) the uppermost 30 cm consists of an alternation from gyttja to peat to gyttja and finally back to peat. This would suggest oscillating water depth with two short-lived intervals of wetter conditions. Liew et al. (1998) show the uppermost 30cm as a soil formed over a thin gyttja layer. We suggest that the presence of gyttja indicates a return to deeper water conditions and that the soil formed within these deposits after the lake dried out at ca 1500 yr B.P. There is insufficient information to determine whether the drying of the lake was due to natural causes or was artificially induced, and we therefore do not code the last 1500 yr of the record.

In the status coding, low (1) is indicated by peat deposition with high loss-on-ignition (50-90%); and high (3) by gyttja with moderate loss-on-ignition (20-50%).



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Liew PM, Kuo CM, Huang SY, Tseng MH (1998) Vegetation change and terrestrial carbon storage in eastern Asia during the Last Glacial Maximum as indicated by a new pollen record from central Taiwan. Global and Planetary Change 16-17: 85-95

Radiocarbon dates

29,300±300	16.80-16.90m, bulk date on peat
28,000±250	15.87-15.97m, bulk date on peat
18,130±160	11.77-11.87m, bulk date on gyttja
12,350±90	9.3-9.41m, bulk date on peat
12,100±90	8.61-8.70m, bulk date on peat
10,450±70	7.89-7.96m, bulk date on peat
9720±60	7.0-7.1m, bulk date on peat
9600±130	7.05-7.07m, AMS date
8780±60	6.1-6.2m, bulk date on peat
8270±70	5.35-5.45m, bulk date on peat
7370±60	4.73-4.83m, bulk date on peat
6480±60	4.2-4.3m, bulk date on peat
5640±60	3.1-3.2m, bulk date on peat
4230±50	1.72-1.82m, bulk date on peat
2230±50	0.8-0.9m, bulk date on peat
1840±50	0.3-0.4m, bulk date on peat

Coding

30,000-25,460 yr B.P.	low (1)
25,460-23,370 yr B.P.	high (3)
25,320-22,410 yr B.P.	low (1)
24,760-20,240 yr B.P.	high (3)
23,500-18,550 yr B.P.	low (1)
18,550-17,620 yr B.P.	high (3)
17,620-11,510 yr B.P.	low (1)
11,920-11,290 yr B.P.	high (3)
11,290-1800 yr B.P.	low (1)
1800-1500 yr B.P.	high (3)
1500-0	not coded

Preliminary coding: 18-3-1999
Final coding: 26-07-2000
Coded by GY and SPH

3.21. Aiding Lake, Xinjiang Autonomous Region

Aiding Lake (42.67°N, 89.27°E, -155m below sea level) is a salt lake in a closed basin in the Tulufan region, inland Xinjing. The basin has very sleep slopes and is surrounded by high mountains (maximum elevation of 5445 m a.s.l.). The change in elevation is 5600m in ca 100km. The Aiding basin originated as a fault depression. Eight rivers terminate in Aiding Lake. The lake water is supplied by runoff from these rivers and some groundwater inflow (Yang et al., 1996). The water in the lake is extremely saline, with a salt content of > 200 g/L (Yang et al., 1996). The lake had an area of 22.5km^2 in 1958 with a mean water depth of 0.8m, but by 1996 the area had been reduced to 5km^2 with a maximum water depth < 0.5m, partly reflecting the long-term trend towards more arid climate. However, the lake area has also been affected by accelerated human activities in the catchment, including the interception of runoff to the lake (Yang et al., 1996). The modern climate in the lake basin is warm and dry, with a mean annual temperature of 14°C, a total annual precipitation of 5mm but evaporation 200-600 times greater than precipitation ($3.189\times10^8 m^3$ per year) (Yang et al., 1996). There is only a sparse vegetation cover of e.g. *Halocnemum. Phragmites* was abundant around the lake until recently, but disappeared due to lake shrinkage and soil salinisation (Yang et al., 1996).

A 51m-long core (Core 86CK1) taken from the salt crust, 500m to the north of the water margin, provides a sedimentary record back to at least the late Pleistocene (Li et al., 1989). Changes in water depth and salinity are reconstructed on the basis of changes in lithology, mineralogy and geochemistry (Li et al., 1989). There are no radiocarbon dates on the sediments below 16.63m. The deposits below 16.63m were thought to have been deposited during the mid- and late- Pleistocene (Li et al., 1989). There are four radiocarbon dates on bulk sediments from the uppermost 16.63m of the core, showing that these deposits span the last 50,000 years.

The basal undated sediments in Core 86CK1 are lacustrine clay (51.01-42.54m), silt (42.54-37.91m), clay (37.91-32.20m), silt (32.20-25.81m), sand (25.81-24.11m), silty clay (24.11-22.0m), sand (22.0-20.0) and silt (20.0-16.63m). Although the lithology of these units changes, suggesting water depth changed significantly, none of the units contain evaporites. Li et al. (1989) therefore interpreted the deposits below 16.63m as indicating a freshwater-lake phase before ca 50,000 yr B.P.

The overlying sediment (16.63-14.11m) is lacustrine silty clay, suggesting moderately deep water. The unit contains 5% evaporites, suggesting the lake water was brackish. A sample from 14.6m was radiocarbon-dated to 39,700±4870 yr B.P. Extrapolation of the sedimentation rate (0.0209 cm/yr) between this date and the overlying radiocarbon date, suggests the bottom boundary of the unit dates to ca 49,420 yr B.P, while interpolation between the two dates indicates that the upper boundary dates to 37,360 yr B.P. Thus, this brackish water phase occurred between ca 49,420-37,360 yr B.P.

The overlying sediment (14.11-11.60m) is silt and sand. Li et al. (1989) suggested this unit was formed by the streams flowing into the lake. This suggests the lake was shallower than before. The unit does not contain evaporites, consistent with freshwater fluvial deposits. The unit is dated to between 37,360-26,900 yr B.P. by interpolation of the sedimentation rate, but may have been formed more quickly than this.

The overlying sediment (11.60-8.36m) is lacustrine silty clay, suggesting increased water depth. The unit contains 20% evaporites (including gypsum and glauberite), suggesting the lake water was brackish. A sample from 11.6m in this unit and a sample from 7.80m in the overlying unit were radiocarbon-dated to 24,900±1240 yr B.P. and 15,700±300 yr B.P. respectively. Thus this brackish-water phase occurred between ca 24,900-17,090 yr B.P.

The overlying sediment (8.36-7.81m) is lacustrine clay, suggesting increased water depth. The unit contains 10% evaporites. The decreased evaporitic content suggests the lake water became fresher, consistent with increased water depth. This unit was dated between ca 17,090-15,720 yr B.P. by interpolation of the sedimentation rate (0.0402 cm/yr).

The overlying sediment (7.81-6.97m) is lacustrine silty clay. The change in lithology indicates decreased water depth. The proportion of evaporites in this unit increased (40%), consistent with the interpretation of decreased water depth. This unit was deposited between ca 15,720-14,330 yr B.P.

The overlying sediment (6.97-6.72m) is lacustrine clay, suggesting increased water depth. The unit contains 10-15% evaporites, consistent with fresher water and increased depth. A sample from the overlying unit (4.90m) was radiocarbon-dated to 10,900±420 yr B.P. Interpolation of the sedimentation rate (0.0604 cm/yr) between this date and the date at 7.80m, indicates this phase of deep water conditions occurred between 14,340-13,170 yr B.P.

The overlying sediment (6.27-4.95m) is lacustrine silty clay, suggesting decreased water depth. The evaporite content increased (20-60%), consistent with shallower conditions. This unit was deposited between 13,170-11,000 yr B.P.

A further decrease in water depth after ca 11,000 yr B.P. is indicated by a layer of gypsum (4.95-4.3m). The evaporite content of the sediments increased to 70%, consistent with relatively shallow conditions. This unit is dated to between 11,000-9560 yr B.P. by interpolation of the sedimentation rate (0.045 cm/yr) between the radiocarbon date from 4.90m and the core top (which is assumed to be modern).

An increase in water depth after ca 8890 yr B.P. is indicated by silty clay (4.3-4.0m) deposition. The decreased content of evaporites (25%) is consistent with relatively fresh and deeper water. This unit is dated to between 9560-8890 yr B.P.

The overlying sediment (4.0-2.93m) is glauberite. The increased evaporite content (70%) is consistent with saline shallow-water conditions. This unit is dated to between 8890-6600 yr B.P.

The overlying sediment (2.97-2.32m) is mirabilite, suggesting a further decrease in water depth. The increased content of evaporites (90%) is consistent with saline, shallow conditions. This unit is dated to between 6600-5160 yr B.P.

An increase in water depth is indicated by a return to glauberite (2.32-0.11m). Li et al. (1989) indicate that the glauberite was formed at a salinity of less than 210 g/L while the mirabilite formed at salinities up to 472 g/L. The evaporite content decreased to 50-60%, consistent with the interpretation of less saline and deeper water conditions. This interval is dated to between 5160-2440 yr B.P.

The overlying sediment (0.11-0.47m) is chlorate and gypsum, suggesting decreased depth. The increased evaporite content (80%) is consistent with more saline and shallower conditions. This unit is dated to between 2440-1040 yr B.P.

The uppermost sediment (0.47-0m) is a salt crust of chlorate and gypsum, indicating the coring site became dry after ca 1040 yr B.P.

In the status coding, extremely low (1) is indicated by salt crust at the coring site; very low (2) by chlorate and gypsum deposition with an evaporite content > 80%; low (3) by mirabilite or gypsum deposition with an evaporite content of 70-90%; moderately low (4) by glauberite deposition with an evaporite content of 50-70%; intermediate (5) by silty clay deposition with an evaporite content of 20-60%; moderately high (6) by clay or silty clay deposition with an evaporite content of <15%; high (7) by fluvial deposition at the coring site during a phase when the lake was freshwater; and very high (8) by clay or silty clay deposition without evaporites.

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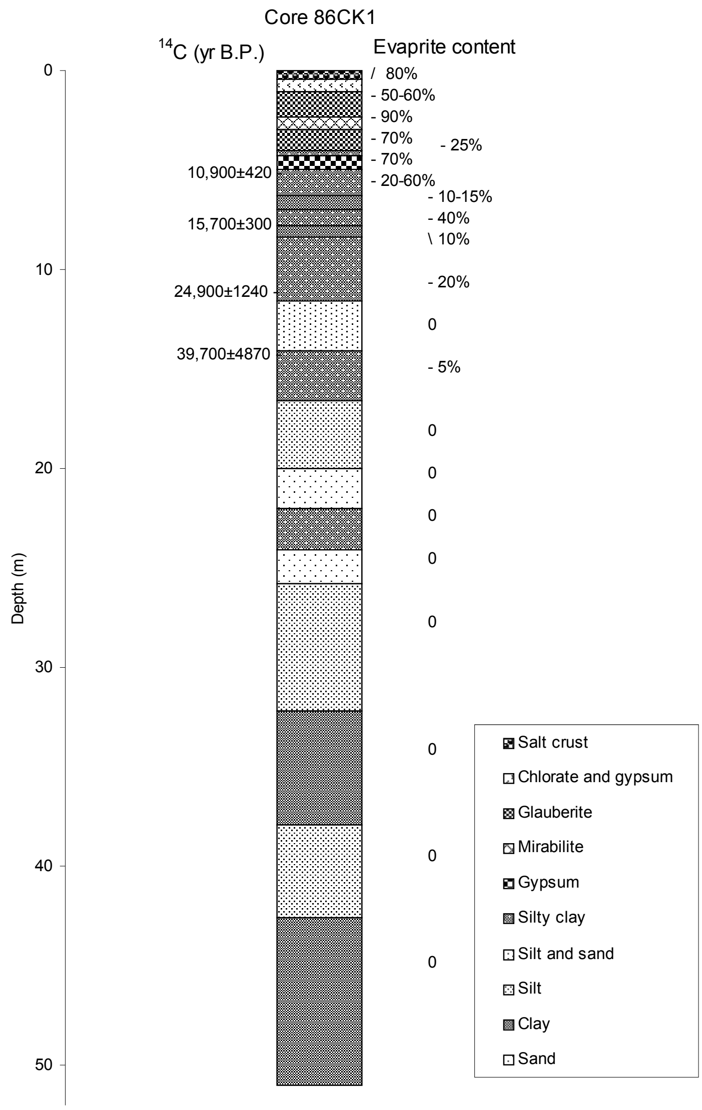
Radiocarbon dates

39,700±4870	14.5m, clay and silt, Core 86CK1
24,900±1240	11.6m, clay and silt, Core 86CK1
15,700±300	7.81m, clay and silt, Core 86CK1
10,900±420	4.95m, clay and silt, Core 86CK1

Coding

pre ca 50,000 yr B.P.	very high (8)
49,420-37,360 yr B.P	moderately high (6)
37,360-24,900 yr B.P.	high (7)
24,900-17,090 yr B.P.	intermediate (5)
17,090-15,720 yr B.P.	moderately high (6)
15,720-14,330 yr B.P.	intermediate (5)
14,330-13,170 yr B.P.	moderately high (6)
13,170-11,000 yr B.P.	intermediate (5)
11,000-9560 yr B.P.	low (3)
9560-8890 yr B.P.	intermediate (5)
8890-6600 yr B.P.	moderately low (4)
6600-5160 yr B.P.	low (3)
5160-2440 yr B.P.	moderately low (4)
2440-1040 yr B.P.	very low (2)
1040-0 yr B.P.	extremely low (1)



Preliminary coding: 21-1-1999
Second coding: 15-3-1999
Coded by GY and SPH

3.22. Aqigekule Lake, Xinjiang Autonomous Region

Aqigekule Lake (in standard Chinese phonetics; given as Aqikol in Huang et al., 1996 and Aqqikkol in the Times World Atlas, 1967) (37.07°N, 88.37°E, 4250m above sea level) is a closed-basin salt lake within the Kumuku Basin, in the middle part of the Kunlun Mountains. The Kumuku Basin is a large structural basin created by faulting. Long-term denudation has subsequently formed a number of inset sub-basins, including the Aqigekule Basin, within the structural basin. The catchment area of Aqigekule Lake itself is unclear, but the Kumuku Basin area is 45,000km^2 (Huang, 1996). The lake area was 395km^2 in 1970 but was reduced to 345km^2 in 1986 because of the long-term trend towards more arid climate (Li, 1992). A few streams enter the lake from the south. The lake water is supplied mainly by runoff and snow meltwater from the basin (Li and Zhang, 1991). The lake has a mean depth of 9.8m (Wang et al., 1987). The water salinity is 78.473 g/L and the pH value is 8.55-8.66 (Li, 1992). The Kumuku Basin is characterised by alpine desert vegetation with Chenopodiaceae and dominated by *Ceratoides* (Huang et al., 1996). The climate in the basin is cold (-5 to -6°C mean annual temperature) and dry (100 to 300mm total annual precipitation) (Huang et al., 1996).

Two lacustrine terraces mark former high stands of the lake. The highest terrace, the top of which is at 4290 m a.s.l. (+40m above modern lake level), occurs 5km southeast of the modern lake shore. A 1.55m-deep sediment profile (Profile D, Li and Zhang, 1991; Li, 1992) was taken in this terrace. A second terrace, the top of which is at 4255 m a.s.l. (+5m above modern lake level), occurs 1km southeast of the modern lake shore. A 5m-deep sediment profile (Profile B given in Li, 1992; Profile E in Li and Zhang, 1991) was taken in this terrace. There is lacustrine lowland to the east of the modern lake, a few meters higher than the modern lake and more than 10km wide. A third profile (Profile East) was taken from these lowlands, ca 8km from the lake and at an elevation of ca 4251-4255 m a.s.l. (Huang et al., 1996). Changes in lake level and lake area are reconstructed from the presence of the preserved shorelines (Li, 1992), and changes in lithology, aquatic pollen assemblages and aquatic plants in the profiles (Li and Zhang, 1991; Li, 1992; Huang et al., 1996) provide a record of changes in relative water depth since ca 16,800 yr B.P. The chronology is based on three radiocarbon dates (Li and Zhang, 1991; Huang et al., 1996).

Profile B from the +5m terrace provides the oldest record from the basin. The basal sediment (5.0-3.70m) is sand with crossbed and ripple strata, and is interpreted as a fluvial deposit (Li and Zhang, 1991). The lake level must have been below 4250 m a.s.l. at this time and was probably lower.

The overlying sediment (3.7-1.3m) is silt and fine sand with thin crossbeds, and is interpreted as a lake-delta deposit (Li and Zhang, 1991). The lake level must have been above 4254 m a.s.l. at this stage and was probably higher.

The overlying sediment (1.3-0.2m) is lacustrine laminated clay, indicating a deep water lake. A sample from the base of the overlying unit (0.15-0.2m) was radiocarbon-dated to 16,765±149 yr B.P. Thus, deep water conditions occurred before ca 16,800 yr B.P. Li (1992) estimated that the high lake level was at ca 4255 m a.s.l. and the lake area was ca 640km^2. Given that the clay deposits are laminated, implying that the lake must have been > 5-10m deep at the profile site, Li's reconstruction must be considered a minimum estimate of both level and area prior to 16,800 yr B.P.

A layer of clay containing abundant aquatic plants occurs between 0.2-0.15m, suggesting decreased water depth after 16,800 yr B.P. The uppermost sediment (0.15-0m) is lacustrine sandy clay, suggesting a further decrease in water depth.

Profile D from the 4290 m a.s.l. terrace provides a mid-Holocene record for the basin. The basal sediments in Profile D (1.55-0.8m) are lacustrine coarse and medium sand, and are interpreted as a beach deposit. At this stage, the lake level must have been ca 4288-4289 m a.s.l. The overlying sediment (0.8-0.2m) is pale-white or pale-yellow lacustrine laminated silty clay to clay, containing fragments of aquatic plants. The lithology and the preservation of laminations suggest deep water. A sample at 0.22-0.38m was radiocarbon-dated to 6705±108 yr B.P. Li (1992) suggests this deep water phase occurred between ca 7000-6500 yr B.P., which seems plausible given the thickness of the unit and assuming a moderate sedimentation rate. Li (1992) estimated that the lake was ca 40m higher than today and 860km^2 in area. Again, given that these deposits are laminated, this must be regarded as a minimum estimate. The uppermost sediment (0.2-0m) is aeolian sand. The lake level had dropped below 4290m after ca 6500 yr B.P.

Profile East, from the lacustrine lowlands to the east of the modern lake provides a record of the late-Holocene deposition in the basin. The basal sediment (below 0.37m) is yellow sand, suggesting shallow water in the basin. The overlying sediment (0.37-0.10m) is lacustrine grey-white laminated silty clay, suggesting increased water depth. The lower part (0.37-0.22m) of the unit is characterised by abundant aquatic plant remains. The upper part (0.22-0.10m) does not contain aquatic remains but *Myriophyllum* is present in the pollen assemblage. A sample from 0.1-0.2m was radiocarbon-dated to 4705±108 yr B.P., suggesting the deep water phase occurred before ca 4700 yr B.P. The top of the lacustrine unit has a similar elevation to the modern lake level, but the site is 8km from the water margin and the sediments are laminated, indicating that the lake was deeper and larger than today. The uppermost sediment (0.1-0m) is beach sand, indicating decreased water depth.

In the status coding, low (1) is indicated by modern lake level at 4250 m a.s.l.; moderately low (2) by lake level of ca 4254-4255 m a.s.l.; intermediate (3) by lake level ca 4260 m a.s.l.; high (4) by lake level ca 4285-4290 m a.s.l.; and very high (5) by lake level higher than 4290 m a.s.l. Although estimates of the minimum lake level can be made for most units from the profiles, there is no means of dating several of these intervals.

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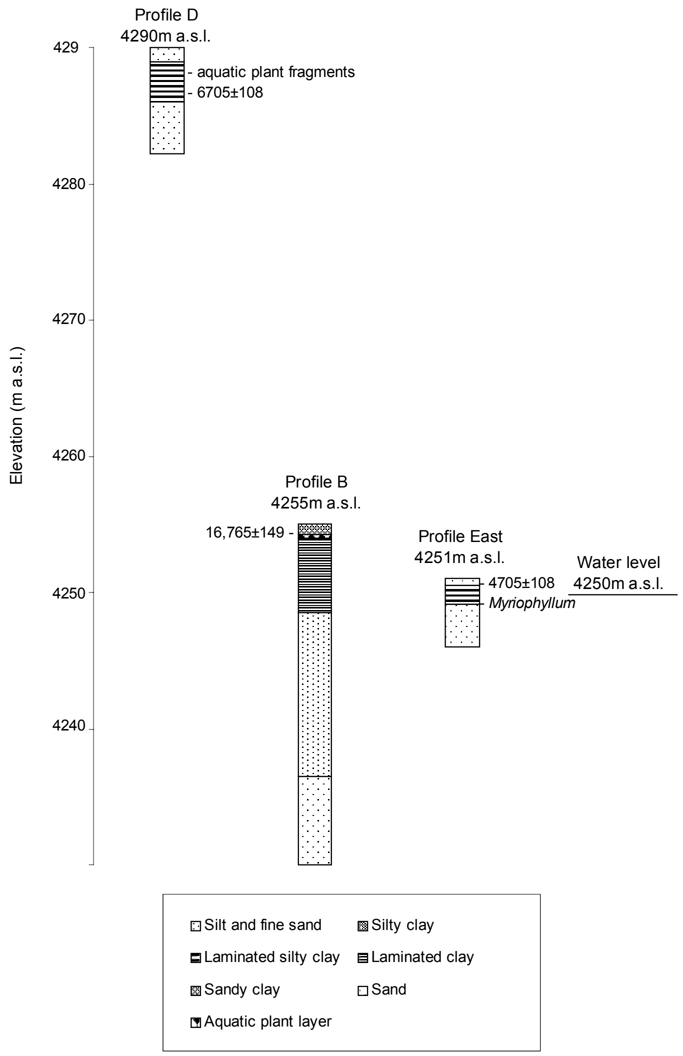


Radiocarbon dates

16,765±149	0.2-0.15m, aquatic plants, Profile B
6705±108	0.22-0.38m, clay, Profile D
4705±108	0.1-0.2m, silty clay, Profile East

(The samples were dated in ^{14}C Lab, Geography Institute of Chinese Academy of Science)

Coding

?	moderately low (2), minimum estimate 4254 m a.s.l.
17,000-16,800 yr B.P.	intermediate (3), minimum estimate 4260 m a.s.l.
?	moderately low (2), ca 4255 m a.s.l.
?	high (4), ca 4288-4289 m a.s.l.
7000-6500 yr B.P.	very high (5), ca 4295-4300 m a.s.l.
5000-4700 yr B.P.	moderately low (2), minimum estimate 4255 m a.s.l.
0 yr B.P.	low (1)

Preliminary coding: 3-12-1998
Final coding: 13-12-1998
Coded by GY and SPH

3.23. Ashikule Lake, Xinjiang Autonomous Region

Ashikule Lake (35.73°N, 81.57°E, 4683m above sea level) is a closed lake in the Ashikule Basin, in the middle part of the Kunlun Mountains, a depopulated zone on the northern Tibetan Plateau (Li, 1992). The Ashikule Basin originated through faulting. Volcanic activity the Quaternary has resulted in the creation of a number of sub-basins which are occupied today by hydrologically-independent lakes (Li and Zhang, 1991). The Ashikule Lake occupies one of these sub-basins. The Ashikule Basin has an area of 740km^2 (Li, 1992), but the catchment area of Ashikule Lake is unknown. A few seasonal streams enter the lake from the north. The bedrock in the lake basin is igneous. The lake water is supplied mainly by runoff and snow meltwater from the basin (Li and Zhang, 1991). The water is saline and the pH value is 9.28 (Li, 1992). The lake area was 11km^2 in 1970, but had been reduced to 10.5km^2 by 1986 due to a long-term trend towards more arid climate (Li, 1992). The climate in the basin is cold (-5 to -6°C mean annual temperature) and dry (100 to 300mm total annual precipitation) (Huang et al., 1996).

Two lacustrine terraces, at 7m (4690 m a.s.l.) and 3-3.5m (ca 4686 m a.s.l.) above modern lake level, provide evidence of former high stands. A 3m-deep sediment profile (Profile A in Li, 1992; Profile E in Li and Zhang, 1991) was taken from the 3 to 3.5m-high terrace, 1km west of the lake margin. A 2m-deep sediment profile (Profile B; Li, 1992) was taken from 7m-high terrace, 3km west of the lake margin. The lake level and area at specific times can be estimated from the terrace shorelines in the basin, while changes in relative water depth between ca 15,800 and 11,600 yr B.P. are reconstructed on the basis of changes in lithology, geochemistry, diatom assemblages and aquatic plants from the sediment profiles (Li and Zhang, 1991; Li, 1992). The chronology is based on two radiocarbon dates from Profile A (Li, 1992).

The basal sediments in Profile A (below 2.3m) are homogeneous lacustrine clays. The unit has relatively high pH values (8.1-8.5) and relatively low organic contents (1.0-5.0%), indicating relatively saline water conditions and no biological activity (Li and Zhang, 1991). This unit indicates relatively shallow water conditions, but nevertheless the lake level must have been > 4684 m a.s.l. (i.e. higher than today). The unit displays disturbed bedding and folded forms, indicating post-depositional modification by permafrost (Li, 1992). This modification probably occurred during the last glacial maximum, and indicates that the lake was not deep at this time.

The overlying sediment (2.3-0.8m) is laminated lacustrine clay, suggesting stable deep water conditions. The unit has pH values of 8.1-7.7 and a relatively high organic content (5.0-7.0%), indicating fresh water and biological activity (Li and Zhang, 1991). The diatom assemblage is dominated by *Cocconeis placentula*, a species tolerant of fresh water (Li JY, 1989; cited in Li and Zhang, 1991). A sample from 2.0m was radiocarbon-dated to 15,256±100 yr B.P. This suggests that the deep water phase occurred between 15,800-13,100 yr B.P.

There is no change in gross lithology in the overlying sediment (0.8-0.08m). However, the unit has higher pH values (8.1-8.4) and lower organic contents (5.0-3.0%), indicating more saline water and less biological activity, suggesting a shallowing after 13,100 yr B.P. The diatom assemblage is characterised by increases in *Nitzschia denticula, Rhopalodia parallela, Rhoicosphenia curvata,* and *Navicula tuscula* which, with the exception of the fresh-water species *Navicula tuscula,* are species tolerant of brackish water. The increase in brackish-water diatoms is consistent with the inference of shallowing indicated by the changes in pH value and organics. This unit is dated to between 13,100-11,780 yr B.P. by interpolation of the sedimentation rate (0.0551 cm/yr) between radiocarbon dates.

The overlying unit (0.08-0.05m) is a layer of clayey aquatic plant fragments. The presence of abundant plant fragments indicates decreased water level. A sample from 0.08-0.05m was radiocarbon-dated to 11,743±260 yr B.P. Extrapolation of the sedimentation rate calculated between the radiocarbon dates suggests this unit formed ca 11,700-11,780 yr B.P.

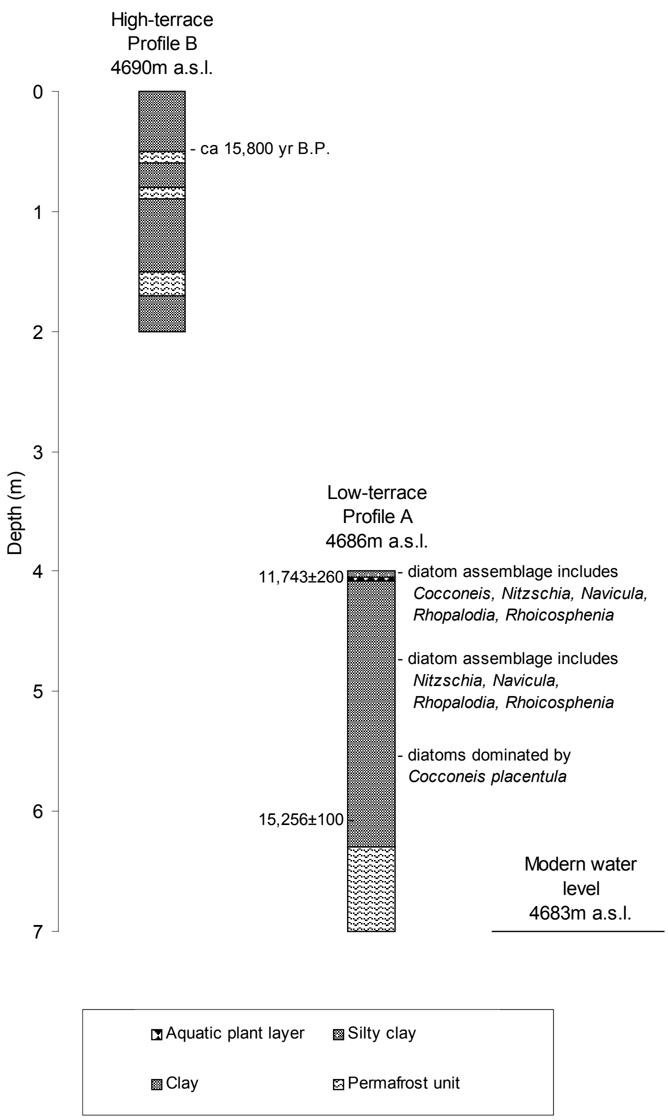
The uppermost sediment (0-0.05m) is lacustrine silty clay, suggesting increased water depth between ca 11,700-11,650 yr B.P. The diatom assemblage is characterised by *Cocconeis placentula*, *Navicula tuscula*, *Nitzschia denticula, Rhopalodia parallela,* and *Rhoicosphenia curvata*. The increase in fresh-water species (*Cocconeis placentula*, *Navicula tuscula*) is consistent with increased depth.

Li (1992) interpreted the 3-3.5m high terrace as representing a lake level at 4686 m a.s.l. around 12,000 yr B.P. At this level, the lake area would have been 16km^2, 5.5 km^2 larger than today. The sediments from Profile A indicate that this lacustrine phase was preceded by an interval, probably corresponding to the glacial maximum, when the lake was shallow and permafrost developed within lacustrine sediments deposited during an earlier lake phase.

Profile B from the 7m-high terrace (4690 m a.s.l.) contains homogeneous lacustrine clay, representing a moderately deep lake. Li (1992) estimated that the lake level ca 7-8m higher than today and the area of the lake was ca 40km^2. Unfortunately the unit is not dated. However, the unit contains three sets of permafrost structures at 0.5-0.6m, 0.8-0.9m and 1.5-1.7m. This suggests that there were three intervals of permafrost activity, alternating with periods of lacustrine deposition. Li (1992) correlated the youngest interval with the permafrost features in Profile A (Li, 1992), which would suggest that the lacustrine deposition occurred before 15,800 yr B.P.

There is no record for the last 11,600 yr B.P. Modern water level is 4683 m a.s.l. (Li, 1992).

In the status coding, low (1) is indicated by modern lake level at 4683 m a.s.l.; moderately low (2) by the lacustrine clayey aquatic plant layer from the 4686 m a.s.l. terrace; intermediate (3) by lacustrine silty clay with brackish and freshwater diatoms in the 4686 m a.s.l. terrace; moderately high (4) by lacustrine laminated clay with pH values above 8.0 and organic values below 5% and a mixed brackish-freshwater diatom assemblage in the 4686 m a.s.l. terrace; high (5) by lacustrine laminated clay with pH values below 8.0 and organic values above 5% in the 4686 m a.s.l. terrace; very high (6) by lacustrine clay from the 7-8m high terrace at 4690 m a.s.l.



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Li SK, Zhang QS (1991) Lake level fluctuations during last 170,000 years in the middle regions of Kunlun Mts. Geographical Research 10(2): 27-37 (in Chinese)

Radiocarbon dates

11,743±260	0.08-0.05m, aquatic plants, Profile A
15,256±100	2.0m, clay, Profile A

(The samples were dated in ^{14}C Lab, Geography Institute of Chinese Academy of Science)

Coding

?	very high (6)
?	low (1), permafrost activity
ca 15,800-13,100 yr B.P.	high (5)
ca 13,100-11,780 yr B.P.	moderately high (4)
ca 11,780-11,700 yr B.P.	moderately low (2)
ca 11,700-11,650 yr B.P.	intermediate (3)
0 yr B.P.	low (1)

Preliminary coding: 04-12-1998
Second coding: 10-12-1998
Final coding: 22-12-1998
Coded by GY and SPH

3.24. Balikun Lake, Xinjiang Autonomous Region

Balikun Lake (43.7°N, 92.8°E, 1575 above sea level according to Han and Dong, 1990; 1585m a.s.l in Li, 1992) is a large salt lake in a closed basin on the northern slope of the Tianshan Mountains, inland Xinjiang. The lake has an area of 112.5 km^2 (Li, 1992) and the catchment area is more than 10,000km^2. The maximum depth of the lake, registered in the southern part of the basin, is between 0.8-1.0m. Permanent snow fields and glaciers occur in the surrounding high mountains, which reach elevation of ca 2200-3800m on the northern side and ca 2500-4300m on the southern side of the basin (Li, 1992). There are more than 60 brooks and rivers draining these high mountains and feeding the Shuimo River, which then supplies inputs to the lake. The lake water is supplied by runoff (ca 50% in total), and snow/glacier meltwater (ca 50%) (Li, 1992). The climate around the lake itself is arid (200-250mm total annual precipitation but 1000-1500mm total annual evaporation) (Li, 1992). Mirabilite is being deposited in the lake today (Han et al., 1993).

The Balikun basin originated as a fault depression and the underlying bedrock is sandstone (Han and Dong, 1990). A lake has been present in the basin since the Early Pleistocene, and there are ca 330m of Quaternary lacustrine sediments in the central part of the basin. Tectonic uplift of the surrounding mountain blocks has persisted through the Quaternary period on timescales of 10^5-10^6 years, and as a result Early, Middle and Late Pleistocene lacustrine sediments are not found around the margin of the basin. Faulting has occurred during the early Holocene, resulting in differential uplift of the oldest Holocene lacustrine terrace (Han and Dong, 1990). According to an internal report of the Xinjiang Geological Survey (1983), evidence for continued tectonism in the late Holocene is provided by an "uplifted accumulative barrier" ca 1.25-2.2m above lake level. Han and Dong (1990) argue, however, that this accumulative barrier is evidence of recent lake-level lowering and not due to tectonic activity. Thus, they consider that the mid- to late-Holocene record from the basin is unaffected by tectonism. There are undated records of glacial advances in the mountains surrounding the Balikun Lake basin assumed to date to the Neoglacial and the Little Ice Age. These advances have been correlated with high lake levels in Balikun Lake, on the assumption that both lake-level changes and glacial advances are controlled by colder and wetter climates. However, there is no evidence that the glacial advances directly influenced lake water balance. Rather, Han and Dong (1990) and Li (1992) assume that the lake-level changes are a direct reflection of climate changes. Thus, although the Balikun Lake record could potentially be affected by tectonic changes and/or by changes in glacial activity driven by changes in climate within the mountains, we have followed the original authors and assumed that these indirect tectonic and/or glacial influences were comparatively unimportant during the last ca 35,000 years and that the lake records can be interpreted as a direct record of changes in the water balance in the basin itself.

A series of lake terraces, which according to the diagram given in Han and Dong (1990) have maximum elevations of 1576m, 1578m, 1585m (T1), 1597m (T2), 1600m (T3), 1604m (T4), 1608m and 1615 m a.s.l. respectively (Han, 1991; Han and Dong, 1990), provide evidence of lake stands higher than present during the Holocene (Han and Dong, 1990). None of the terraces have been independently radiocarbon-dated, although Han and Dong (1990) have attributed ages to the terraces on the basis of correlation with the implied climatic changes recorded in lacustrine core records from the basin. There are discrepancies in the published paper in the terrace names, elevations and possible ages. It is therefore impossible to use the terrace information to code lake-level changes. We summarise the available information here, but make no use of it for coding purposes.

The two highest terraces (T1615 and T1608) are thought to be relicts of a single terrace, deformed by subsequent tectonism (Han and Dong, 1990). T1615 has basal deposits of pebble-gravels, overlain by lacustrine clay and with sand at the top (Han, 1991). Terrace T4 (1604m) consists of lacustrine clay with sandy clay on the top (Han, 1991). According to Han and Dong (1990) this terrace was formed between 7000-8000 yr B.P., and the associated lake would have been ca 29m higher than the modern lake. Terrace T3 (1600m) consists of pale-grey lacustrine silty clay (Han, 1991). According to Han and Dong (1990) this terrace was formed between 6000-5000 yr B.P., and the associated lake would have been ca 25m higher than the modern lake. Terrace T2 (1597m) consists of brown-black lacustrine clay with a layer of sand on the top (Han, 1991). According to Han and Dong (1990) this terrace was formed ca 3000 yr B.P., and the associated lake would have been ca 22m higher than the modern lake. Terrace T1 (1585m) consists of pale-grey lacustrine silty clay with a layer of sand on the top (Han, 1991). According to Han and Dong (1990) the terrace was formed ca 1700 yr B.P., and the associated lake would have been ca 10m higher than the modern lake. There are two lower, unnamed terraces, at ca 1578m and 1576m. According to Han and Dong (1990) the higher of these terraces was formed between 1700 and 100 yr B.P. and the lower of these terraces ca 100 to 50 yr B.P. The present lake level (1575m) is the lowest during the Holocene (Han and Dong, 1990).

The information about the lacustrine terraces can be summarised as follows:

Terrace name (from diagram)	Terrace name (from text)	Elevation (m above modern lake level) as given in text	Maximum terrace elevation (m a.s.l.), measured from diagram	Elevation (m a.s.l.) of terrace surface, listed on diagram	Implied lake level, above modern (m)	Estimated age (yr B.P.)
T1615			1615	-	40	oldest
T1608			1608	1608	33	same as T1615, elevation difference due to faulting
T4			1604	1604	29	7000-8000
T3	t2-2	17-1.5m (1592-1576.5 m a.s.l.)	1600	1600	25	6000-5000
T2	t2-1	2.2-1.8 (1577.2-1576.8 m a.s.l.)	1597	1597	22	3000
T1	t1	1.25 (1576.25m a.s.l)	1585	-	10	1700
unnamed			1578	1564 (error)	3	1700-100
unnamed			1576	-	1	100-50
modern lake			1575	-	0	0

A more reliable record of relative lake-level changes in the Balikun basin is provided by two lacustrine cores. A 13.6m-long core (Core ZK00A) taken from the northwest of the lake provides a sedimentary record back to ca 37,000 yr B.P. (Han and Yuan, 1990). A second 10.76m-long core (Core ZK0024) from the centre of the lake basin provides a record back to ca 25,000 yr B.P. (Han and Dong, 1990). The two core records can be correlated on the basis of their lithology (Han et al., 1993). There is an ostracode record from Core ZK00A (Han et al., 1993). Unfortunately, the summary diagram in Han et al. (1993) only shows two species (*Eucypris inflata* and *Leucocythere mirabilis*), and ostracodes are absent or in low abundance in many parts of the core. It is difficult to reconcile the changes in the presence/abundance of ostracodes with the lithological record and it appears that some of these changes may represent post-depositional modification of the assemblages e.g. through dissolution in highly saline water. Although we summarise the ostracode record, our interpretations of changes in lake-levels are therefore based entirely on changes in lithology and geochemistry. According to Han et al. (1993) a total of 43 samples were taken for radiocarbon dating from Core ZK00A; 7 of the samples were dated by the Archaeology Institute of the Chinese Academy of Science and no information is available about these dates. The remaining 36 samples were dated by Guangzhou Geography Institute and are used in Han and Yuan (1990) and Han et al. (1993). Unfortunately, only 26 of these dates are documented in the Han et al. (1993) paper. The additional dates appear to be shown in the diagrams in Han and Yuan (1990); however, there are multiple discrepancies between the dates given in various parts of this paper and between this paper and the dates listed in Han et al. (1993). It is therefore not possible to reconstruct a complete and accurate date list for Core ZK00A. We therefore construct the chronology for Core ZK00A using only the 26 fully-documented dates from Han et al. (1993). There are two additional radiocarbon dates from Core ZK0024 (Han and Dong, 1990); these two dates are also used in erecting the core chronology. All of the dates (including dates which we are unable to harmonise) are listed in the radiocarbon-date list below.

The basal sediment from core ZK00A (13.6-13.0m) is grey and brown clay with fine silt. The unit is of lacustrine origin, and suggests a relatively deep-water lake was present in the basin. There are no ostracodes present in the basal part of the unit, but a single sample at ca 13.2m contains low amounts of *Eucypris inflata* (a species indicating relatively saline conditions). It is unclear how this single ostracode record can be interpreted. A sample at 13.6m was radiocarbon-dated to 36,700±829 yr B.P. and a sample at 13.30m was dated to 35,100±740 yr B.P. Using the sedimentation rate between these two dates suggests that the end of the deep-water phase occurred ca 33,500 yr B.P. A radiocarbon date spanning the boundary gives an age of 33,710±170 yr B.P, consistent with the estimated age. Thus, we assume that the initial, relatively deep-water phase occurred between ca 37,000 and 33,600 yr B.P.

The overlying sediment is calcareous clay (13.0-12.70m). The increased carbonate content suggests a decrease in water depth. There is no ostracode record from this unit. However, the presence of aquatic pollen (species not given) is consistent with decreased water depth. A sample from the base of the unit (13.0m) was radiocarbon-dated to 33,710±170 yr B.P. and a sample at the top of the unit (12.70m) was dated to 32,850±670 yr B.P. This suggests the shallower-water phase occurred between 33,600 and 32,850 yr B.P.

The overlying sediment (12.70-12.10m) is grey-green silty clay. The change in lithology is consistent with increased water depth. A decrease in the abundance of aquatic pollen is consistent with this interpretation. The ostracode assemblage is characterized by low abundances of *Eucypris inflata*, as in the case of the basal, non-calcareous silt-clay unit. A sample at 12.10cm was radiocarbon-dated to 31,950±110 yr B.P. Interpolating the sedimentation rate (0.041 cm/yr) between this date and a date from the overlying unit indicates that this deeper-water interval occurred between ca 32,850 and 32,000 yr B.P.

The overlying sediment (12.10-11.50m) is silty clay with gravel and fine pebbles. The presence of gravels and pebbles is consistent with shallowing. There is no significant change in the ostracode record, which is still characterised by the continuous presence of *Eucypris inflata* in low abundance. A sample from the top of the unit (11.50m) was dated to 30,490 yr B.P. This suggests a decrease in water depth, ca 32,000-30,500 yr B.P.

The overlying sediment (11.50-7.90m in Core ZK00A, ca 10.76-9.62m in Core ZK0024) is lacustrine clay. The colour in the deeper-water core (ZK0024) is darker than in the shallower-water core (ZK00A), where it is described as a grey and yellow lacustrine clay. The change in lithology indicates a significant increase in water depth. The top and bottom of this unit in ZK00A are radiocarbon-dated to 30,490 and 24,100 yr B.P. respectively. A sample from near the top of the unit in Core ZK0024 was radiocarbon-dated to 24,310±225 yr B.P., and thus suggests a similar timing for the end of this relatively deep-water phase. There is no ostracode record for this unit from Core ZK0024. The ostracode record from Core ZK00A initially shows no change in the ostracode assemblage, being characterised by continued low abundance of *Eucypris inflata*. A single sample at ca 10.5m is characterised by the presence of *Eucypris inflata* and high abundance of *Leucocythere mirabilis* (ca 10.5-10.0m), a species that occurs in less saline conditions than *Eucypris inflata.* In the subsequent sample, however, *Leucocythere mirabilis* declines and *Eucypris inflata* increases markedly in abundance. Between ca 10.1m and 9.3m, *Eucypris inflata* is present in low abundance and *Leucocythere mirabilis* is absent. The uppermost part of the unit is characterised by high abundances of *Eucypris inflata* and the continued absence of *Leucocythere mirabilis*. At face value, the ostracode record shows an initial moderately saline phase between 30,500 to 27,700 yr B.P., a relatively freshwater phase at ca 27,700 yr B.P., a highly saline phase at ca 27,350 yr B.P., a moderately saline phase from 27,250 to 26,830 yr B.P., and a return to a highly saline phase between 26,830 and 24,100 yr B.P. However, these changes are not supported by the lithological/geochemical record from the core and are difficult to interpret.

The overlying unit in Core ZK00A (7.90-6.42m) is a grey and yellow fine sandy clay. The change in lithology suggests a decrease in water depth. The ostracode record shows an initial phase with low *Eucypris inflata* (7.90-7.00m) followed by a phase of high abundance of *Eucypris inflata* (7.00-6.42m). Decreased water depth is also registered by a shift to silty clay (ca 9.62-9.2m) in Core ZK0024. The difference in lithology is consistent with the fact that ZK0024 is from a more central (i.e. deeper water) location than Core ZK00A. The sediments for ZK0024 suggest that the shallow water phase was interrupted by a minor oscillation towards deeper-water conditions, represented by deposition of clay between ca 9.2-8.6m and then a return towards shallow conditions, represented by the deposition of clayey silt between ca 8.6-8.22m. Samples from the bottom (ca 7.9m) and top (ca 6.42m) of the sandy clay unit in Core ZK002 were radiocarbon-dated to 24,100±? and 20,730±500 yr B.P. respectively. Independent estimates of the length of this interval based on the two dates from Core ZK0024 are broadly consistent with this interpretation, suggesting the interval occurred between 23,500 and 20,160 yr B.P. On the basis of these dates, the interval of deeper water conditions which interrupted this shallow-water phase as registered by clay deposition in Core ZK0024 occurred ca 22,500-20,700 yr B.P. Given the paucity of dates on Core ZK0024, and the several-hundred year discrepancy in the timing of the shallow-water interval as estimated from this core and Core ZK0024, it is not possible to code the oscillations within the shallow-water phase as independent units.

The overlying unit (6.42-5.60m in Core ZK00A, 8.22-7.84m in Core ZK0024) is laminated lacustrine clay. In Core ZK00A, the unit is described as a grey-white clay with thin banding, and corresponds to an interval with high calcium carbonate content. The lithology and geochemistry suggest that the lake was relatively shallow, but given the absence of sand-sized material, must have been somewhat deeper than before. The presence of aquatic pollen is consistent with this interpretation. The ostracode record shows very high abundance of *Eucypris inflata*, which is consistent with shallow and saline conditions. This interval of moderately shallow conditions is radiocarbon-dated to between 20,730±500 yr B.P. and 17,800±470 yr B.P.

The overlying sediment (5.60-4.20m in Core ZK00A, ca 7.84-7.48m in Core ZK0024) is grey and yellow lacustrine clay. The change in colour, and the absence of banding, is consistent with increased water depth after ca 17,800 yr B.P. The basal part of the unit (>5m) in Core ZK00A is characterised by extremely abundant *Eucypris inflata* but the upper part of the unit (4.20-5.0m) shows moderate to high abundance of *Leucocythere mirabilis*, and this change may be consistent with freshening and deeper water.

The overlying sediment (4.20-2.62 m in Core ZK00A, 7.48-6.80m in Core ZK0024) is silty clay. In the more central core (ZK0024) the sediment contains carbonate nodules, while in the shallower-water core (ZK00A) the sediment is rather organic (as indicated by its grey-black colour). The presence of carbonate and of organics is consistent with the interpretation of shallowing. The ostracode assemblage is characterised by low abundance of *Leucocythere mirabilis*, and it is possible that the decreased abundance of this unit (and its ultimate disappearance) compared to the underlying unit is consistent with shallowing. Samples from the bottom (4.2m) and top (2.62m) of the unit have been radiocarbon-dated to 14,360±410 yr B.P. and 12,150±240 yr B.P. respectively.

The overlying sediment (2.62-1.75 m in Core ZK00A, 6.80-6.62m in Core ZK0024) is white-grey mirabilite mud or clay. The presence of mirabilite indicates considerably increased salinity and very much shallower conditions after ca 12,150 yr B.P. A series of radiocarbon dates indicate that mirabilite formation persisted at the site of Core ZK00A until ca 8450 yr B.P. However, the thinness of the mirabilite-rich unit in the more central part of the lake suggests that the maximum shallowing was registered during the early part of this period. During the latter part of the interval that mirabilite was being deposited in Core ZK00A, silty clay (6.62-5.83m) and clayey silt containing abundant plant remains (5.83-5.60m) was deposited in Core ZK0024. This sequence suggests that water depth increased somewhat, although the abundance of plant remains indicates that the increase was small. Estimates of the timing of these changes based on the two radiocarbon dates on Core ZK0024 yield ages for the beginning of mirabilite deposition and the termination of shallower conditions that are ca 4000 yr different from those derived from the better-dated core ZK00A core. Estimates based on using the dates for the beginning and end of the mirabilite deposition phase in ZK00A and assuming that these correspond to the beginning and end of the shallow-water phase in ZK0024 suggest that the initial pronounced shallow water phase (represented by mirabilite formation in ZK0024) occurred ca 12,150 to 11,550, the somewhat deeper water phase (represented by organic silty clay deposition) occurred ca 11,550 to 9160, and the final shallow-water phase (represented by the plant-rich layer) occurred ca 9160 to 8450 yr B.P.

The overlying sediment (1.75-1.50 m in Core ZK00A, 5.60-2.90m in Core ZK0024) is lacustrine mud. The change in lithology indicate a return to fresh-water conditions and an increase in water depth. The presence of *Candoniella leatea* (5.60-2.90m*)*, an ostracode characteristic of fresh to brackish water, in the central core ZK0024 is consistent with freshening. Based on the radiocarbon dates from Core ZK00A, this freshwater interval dates to between ca 8450 and 7000 yr B.P. A sample from the top of the unit (2.90m) in Core ZK0024 was radiocarbon-dated to 7495±65 yr B.P., suggesting a timing for the end of the phase broadly consistent with that based on Core ZK00A.

The overlying sediment (1.50-1.20 m in Core ZK00A, 2.9-1.5m in Core ZK0024) is mirabilite mud. The change in lithology indicate increased salinity and a return to shallow-water conditions. The disappearance of *Candoniella leatea* within the mirabilite mud from Core ZK0024 (above 2.90m) is consistent with decreased water depth. Based on the radiocarbon dates from Core ZK00A, this interval occurred between 7000-6500 yr B.P.

The overlying unit (1.20-0.78 m in Core ZK00A, 1.5-0.8m in Core ZK0024) is black-grey lake mud and contains no mirabilite. The change in lithology indicates that the lake became fresh and moderately deep. The ostracode assemblage from Core ZK0024 is characterized by the return of *Candoniella leatea,* consistent with freshening and an increase in lake depth. A sample from the upper boundary (0.78 m) of this unit in Core ZK00A was radiocarbon dated to 4130±116 yr B.P.

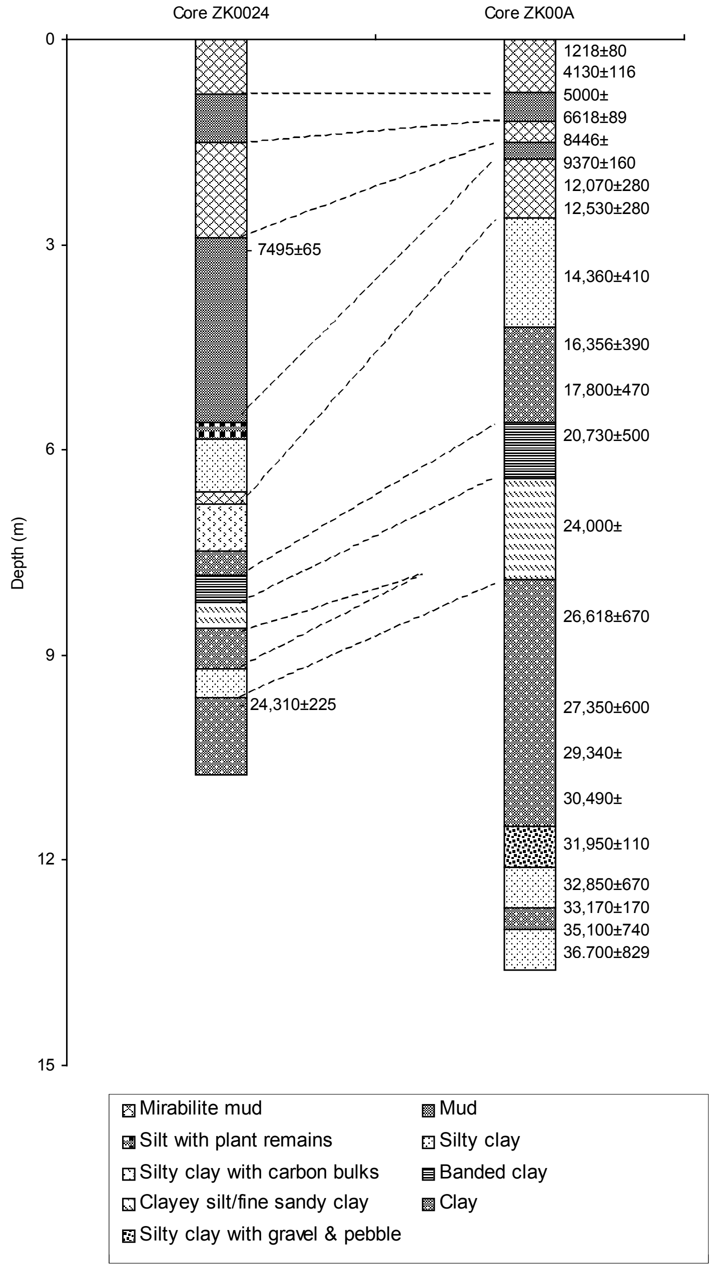
The uppermost sediment (0.78-0.02 m in Core ZK00A, 0.8-0.0m in Core ZK0024) is mirabilite mud, indicating increased salinity and shallowing after ca 4130 yr B.P. A sample from 0.36 m in Core ZK00A was radiocarbon dated to 1218±80 yr B.P., confirming that this saline-phase has persisted for several thousand years. The coring site of ZK00A is now dry and there is a thin saline crust at the top of the core.

In the status coding, extremely low (1) is indicated by mirabilite mud in both cores; very low (2) is indicated by mirabilite deposition in Core ZK00A and shallow-water deposits with abundant plant remains in Core ZK0024; low (3) by mirabilite deposition in Core ZK00A and organic silty clay deposition in Core ZK0024; moderately low (4) by silty clay with gravels or clays with sands; intermediate (5) by calcareous clay deposition in Core ZK00A; moderately high (6) by banded calcareous clays in both cores; high (7) by silt clays, rich in organics in Core ZK00A and with carbonate nodules in Core ZK0024; very high (8) by silty clays or clays containing silt; and extremely high (9) by lacustrine clays.

Radiocarbon dates

36,700±829	13.60m, clay, in Core ZK00A, Ref 3
35,100±740	13.30m, silty clay, in Core ZK00A, Ref 3, Ref 2
33,710±170	13.00m, calcareous clay, in Core ZK00A, Ref 3 (given as 33,780 in Ref 2)
32,850±670	12.70m, calcareous clay, in Core ZK00A, Ref 3 (given as 32,840 in Ref 2)
31,950±110	12.10m, silty clay, Core ZK00A, Ref 3
31,432±	Core ZK00A, Ref 2
30,490±	11.50m, silty clay, Core ZK00A, Ref 3, Ref 2
29,470±	11.50m, silty clay, Core ZK00A, Ref 3, Ref 2
29,340±	11.10m, silty clay, Core ZK00A, Ref 3 (given as 29,342 in Ref 2)
28,847±	Core ZK00A, Ref 2 (also given as 28,849 on Fig 2)
27,350±600	10.30m, clay, Core ZK00A, Ref 3 (given as 27,370 in Ref 2)
26,618±670	8.90m, silt and clay, Core ZK00A, Ref 3
26,413±	Core ZK00A, Ref 2 (also given as 26,410)
25,670±	Core ZK00A, Ref 2
24,489±	Core ZK00A, Ref 2
24,310±225	ca 9.9m, clay, Core ZK0024, Ref 1
24,100±	7.90m, clay, Core ZK00A, Ref 3
23,120±	Core ZK00A, Ref 2
21,774±	Core ZK00A, Ref 2 (also given as 21,770)
20,730±500	6.42m, sandy clay, Core ZK00A, Ref 3
20,200±	Core ZK00A, Ref 2
17,800±470	5.60m, laminated clay, Core ZK00A, Ref 3 (given as 17,860 in Ref 2)
16,356±390	5.00m, silty mud, Core ZK00A, Ref 3
16,176±360	5.00m, silty mud, Core ZK00A, Ref 3
15,829±	Core ZK00A, Ref 2
15,170±	Core ZK00A, Ref 2
14,360±410	4.20m, clay, Core ZK00A (given as 1436±410 in text), Ref 3, Ref 2
13,978±	Core ZK00A, Ref 2
13,814±	Core ZK00A, Ref 2
12,747±	Core ZK00A, Ref 2
12,530±	2.84m, silty clay, Core ZK00A, Ref 3, Ref 2
12,150±240	2.62m, silty clay, Core ZK00A, Ref 3
12,070±280	2.51m, mirabilite clay, Core ZK00A, Ref 3
10,870±	Core ZK00A, Ref 2
10,084±	2.20m, clayey mirabilite, Core ZK00A, Ref 3, Ref 2
9370±160	2.00m, clayey mirabilite, Core ZK00A, Ref 3
8970±	Core ZK00A, Ref 2
8446±160	1.70m, mirabilite clay, Core ZK00A, Ref 3
8190±	Core ZK00A, Ref 2
7495±65	ca 2.90m, mud, Core ZK0024 (given as 7486±65 and 7495±250 in text), Ref 1
6618±89	1.20-1.25m, mud, Core ZK00A (given as 6668±89 in text), Ref 3, Ref 2
5000±	1.04m, clay, Core ZK00A, Ref 3
4130±116	0.78m, mud, Core ZK00A, Ref 3 (given as 4180 in Ref 2)
3270±	Core ZK00A, Ref 2
2640±	Core ZK00A, Ref 2
2370±	Core ZK00A, Ref 2
2310±	Core ZK00A, Ref 2
1700±65	date associated with accumulative barrier, Ref 1
1218±80	0.02-0.36 m, mirabilite, Core ZK00A, Ref 3

Ref 1: Han and Dong (1990); Ref 2: Han and Yuan (1990); Ref 3: Han et al. (1993)



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Coding

37,000-33,600 yr B.P.	very high (8)
33,600-32,850 yr B.P.	intermediate (5)
32,850-32,000 yr B.P.	very high (8)
32,000-30,500 yr B.P.	moderately low (4)
30,500-24,100 yr B.P.	extremely high (9)
24,100-20,730 yr B.P.	moderately low (4)
20,730-17,800 yr B.P.	moderately high (6)
17,800-14,360 yr B.P.	extremely high (9)
14,360-12,150 yr B.P.	high (7)
12,150-11,550 yr B.P.	extremely low (1)
11,550-9160 yr B.P.	low (3)
9160-8450 yr B.P.	very low (2)
8450-7000 yr B.P.	extremely high (9)
7000-6500 yr B.P.	extremely low (1)
6500-4130 yr B.P.	extremely high (9)
4130-0 yr B.P.	extremely low (1)

Preliminary coding: 10-11-1998
Second coding: 20-11-1998
Third coding: Nov 2000
Final coding: 21-01-2001
Coded by GY and SPH

3.25. Beilikekule Lake, Xinjiang Autonomous Region

Beilikekule Lake (in standard Chinese phonetics; given as Beilikol in Huang et al., 1996 and Berikekule in Li, 1992) (36.72°N, 89.05°E, 4680m above sea level) is a closed-basin lake in the Kumuku Basin, in the middle part of the Kunlun Mountains. The Kumuku Basin is a large structural basin; long-term denudation has created a number of sub-basins, including Beilikekule basin, within this structural basin. The catchment area for Beilikekule Lake is unclear, but the Kumuku Basin area is 45,000km^2 (Huang, 1996). Two seasonal streams enter Beilikekule Lake from the southwest. The lake water is supplied mainly by runoff and snow meltwater from the basin (Li and Zhang, 1991). The lake area was 5.0km^2 in 1970 and was reduced to 4.4km^2 by 1986 due to the long-term trend towards more arid climate (Li, 1992). The water salinity is 2.523 g/L and the pH value is 9.53 (Li, 1992). The Kumuku Basin is characterised by alpine desert vegetation with Chenopodiaceae and dominated by *Ceratoides* (Huang et al., 1996). The climate in the basin is cold (-5 to -6°C mean annual temperature) and dry (100 to 300mm total annual precipitation) (Huang et al., 1996).

Two lacustrine terraces, one 25m and the other 2-3m above modern lake level, indicate former high stands of the lake. A 6m-deep sediment profile (Profile C in Li and Zhang, 1991; called Profile F in Li, 1992) was taken from the 25m terrace (the top of which is at 4705 m a.s.l., +25m above modern lake level), 1.5km southeast of the lake margin. The profile provides a sedimentary record back to ca 12,400 yr B.P. The preserved shorelines are used as a basis to estimate palaeolake areas at specific times (Li, 1992). Changes in relative water depth are reconstructed on the basis of changes in lithology, aquatic pollen assemblages and aquatic plants (Li and Zhang, 1991; Li, 1992) shown in Profile C. The chronology is based on two radiocarbon dates from Profile C (Li and Zhang, 1991).

The basal sediments in Profile C (5.0-6.0m) are silt and fine sand with crossbed and ripple strata, interpreted as a fluvial deposit (Li and Zhang, 1991). The lake level must have been below 4699m at this time and was likely much lower.

The overlying sediments (5.0-4.5m) are lacustrine clay containing abundant aquatic plant material. The change in lithology suggests increased water depth. The aquatic pollen assemblage is characterised by *Myriophyllum* and *Scheuchzeria*, consistent with moderately deep water at the site. A sample from 4.75m is radiocarbon-date to 12,253±280 yr B.P. If we use sedimentation rate between the date from and the date of 6311±77 yr B.P. from 1.6m from a second lacustrine unit in this profile, then this lacustrine phase this unit is dated to 12,735-11,750 yr B.P. However, the two radiocarbon samples on lacustrine units are separated by a layer of alluvial sand, and interpolation between the dates on the assumption that the sedimentation rate was constant may not be justified.

Li (1992) assumed that this lacustrine unit, the top of which is at an elevation of ca 4700 m a.s.l., represented a high lake stand, ca 20m above the modern lake level. He estimated the lake area defined by the 4700m contours in the basin to be 18km^2. However, there are no shorelines in the basin at 4700m (only 4705m terrace). Furthermore, the clay unit was likely to have been formed in a water depth of at least 0.5 to 1m (i.e. lake level of ca 4701m). Thus, we believe that the lake must have been larger than 18km^2 estimated by Li (1992) at this time.

The overlying sediment (4.5-2.0m) is pale-white silt and fine sand. It does not contain aquatic plants or organics. This was interpreted as an alluvial deposit (Li, 1992). The lake level must again have fallen below ca 4700 m a.s.l.

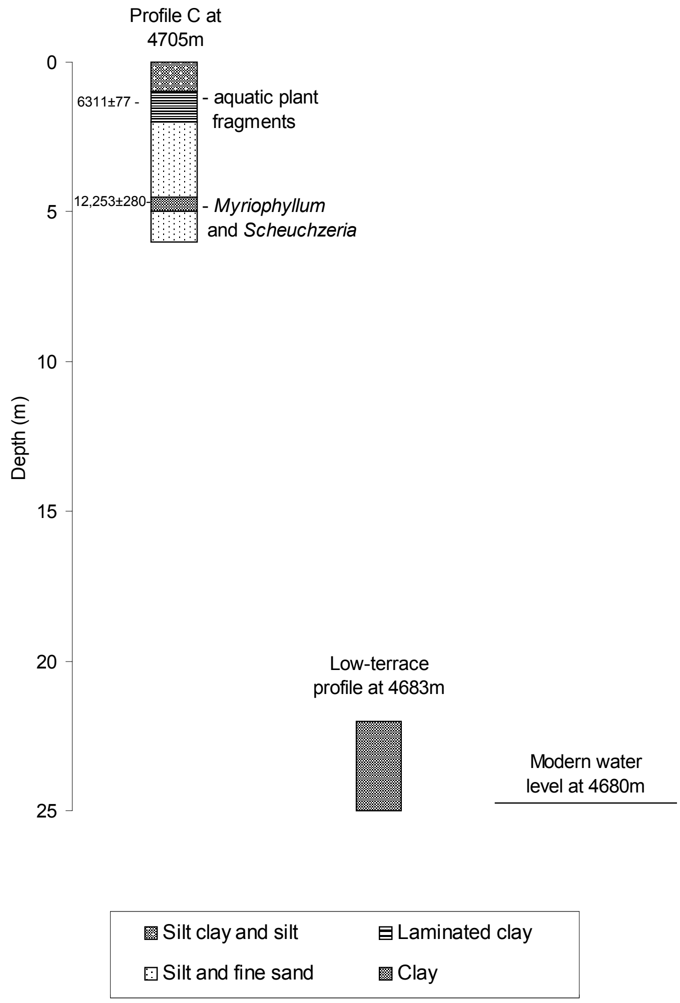
The overlying sediment (2.0-1.0m) is laminated lacustrine clay containing fragments of aquatic plants, suggesting relatively deep water conditions at the site. A sample from 1.6m is radiocarbon-dated to 6311±77 yr B.P. Li and Zhang (1991) and Li (1992) suggest that the unit was formed between ca 7000 and 6000 yr B.P., which seems plausible assuming normal sedimentation rates for lacustrine clays. Huang et al. (1996) correlated this phase, which is characterised by high percentages of *Artemisia* and low percentages of Chenopodiaceae, with a similar pollen assemblage in the nearby site Aqigekule, and also suggest, on the basis of two mid-Holocene radiocarbon dates from Aqiugekule, that the laminated clay unit at Beilikekule was formed between ca 7000 and 6000 yr B.P. However, we note that the radiocarbon dates from Aqigekule are from two different lacustrine phases, apparently separated by a somewhat drier interval marked by aeolian deposition. Therefore, the chronology from Aqigekule may not be sufficiently reliable to transfer in this way. Linear interpolation using the sedimentation rate (0.053 cm/yr) between the two radiocarbon dates from this profile gave the upper and lower boundaries between 5180-7065 yr B.P., but this could be an overestimate as the fluvial deposit between the two radiocarbon-dated; lacustrine beds may have been deposited quite rapidly.

The uppermost sediment (0-1.0m) is lacustrine silty clay and silt, interpreted as a nearshore deposit (Li and Zhang, 1991). This unit indicates that lake level was at or slightly above 4705m shortly after 6000 yr B.P. The cessation of deposition indicates that the lake fell below 4705m. The upper boundary is estimated to ca 5500 yr B.P. by pollen correlation with Aqigekule record (Huang et al., 1996) and to ca 3300 yr B.P. by extrapolation of the sedimentation rate between two radiocarbon dates.

Li (1992) suggested that the laminated unit, with an upper elevation of ca 4704m, represents a high lake stand, ca 24m above the modern lake level and roughly estimated the lake area as ca 26km^2 on the basis of the intermittent shorelines along the ca 4705m contours in the basin. He suggested this 26 km^2 large lake was present between 7000-6000 yr B.P. (Li, 1992). We argue that the water level during the interval when the laminated unit was formed must have been considerably higher than 4704-4705 m a.s.l. since laminations are not preserved in water only 1m deep. Therefore the lake area was probably larger than 26km^2 between 7000-6000 yr B.P. The overlying nearshore deposit, however, must have been formed when the water depth was close to 4705m. Thus, Li (1992)'s estimate of 26 km^2 is more closely applicable to the size of the lake when the uppermost unit was formed.

There is a low-terrace, 2-3m above modern lake level around the lake. The well preserved form of this terrace indicates that it is comparatively young. Li and Zhang (1991) suggest that it marks a 2-3m high lake stand during the last 1200 years. The position of these terraces clearly indicates that they are younger than 5500 yr B.P. Although the age of 1200 yr B.P. is plausible, there is no compelling reason to assume that the terraces were not formed earlier.

In the status coding, low (1) is indicated by fluvial deposition in the 25m-terrace, by modern lake level, or by the low terrace 2-3m above modern lake level; moderately low (2) by nearshore deposits of silty clay and silt from the 25m-terrace; intermediate (3) by lacustrine clay with abundant aquatic plants from the 25m-terrace; high (4) by lacustrine laminated clay with aquatic plant fragments from the 25m-terrace.



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Radiocarbon dates

12,253±280	4.75m, aquatic plants, Profile C
6311±77	1.6m, clay, Profile C

(The samples were dated in ^{14}C Lab, Geography Institute of Chinese Academy of Science)

Coding

pre- ca 12,735 yr B.P.	low (1)
ca 12,735-11,750 yr B.P.	intermediate (3)
ca 11,750-7065 yr B.P.	low (1)
ca 7065-5180 yr B.P.	high (4)
ca 5180-3290 yr B.P.	moderately low (2)
ca 3290-0 yr B.P.	low (1)

Preliminary coding: 1-12-1998
Second coding: 18-12-1998
Final coding: 20-1-1999
Coded by GY and SPH

3.26. Chaiwopu Lake, Xinjiang Autonomous Region

Chaiwopu Lake (43.5°N, 87.9°E, 1092m above sea level) is a closed salt lake in the Chaiwopu Basin on the northern slope of the Tianshan Mountains, inland Xinjiang. The basin originated through faulting. The bedrock in the basin consists of Carboniferous-Permian detrital rocks, with some carbonate and igneous rocks (Zhang and Wen, 1990). Chaiwopu Lake has an area of 29km^2. The lake has a mean depth of 4.22m and a maximum depth of 6.10m. The lake water is supplied by runoff and snow/glacier meltwater from the high mountains (maximum elevation of 5445 m a.s.l.) around the basin. The catchment area is ca 400km^2. The mean annual temperature in the basin is 8.6 °C, the total precipitation is 50mm and the annual evaporation is 3365.5mm. In the surrounding mountains (above 4000 m a.s.l.) the mean annual temperature is -5.5°C, and total precipitation rises to 700-750mm while annual evaporation is reduced to 140mm (Wang and Jiao, 1989). The Chaiwopu basin is characterised by warm desert vegetation, dominated by Chenopodiaceae (including *Haloxylon, Anabasis, Nanophyton* and *Ceratoides*), *Ephedra, Zygophyllum* and Tamaricaceae (including *Reaumuria*) (Li and Yan, 1990).

Discontinuous erosional terraces consisting of lacustrine sediments have been identified at 3 elevations above the modern lake level. The T1 terrace is ca 1-2m, the T2 terrace is ca 7-8m and the T3 terrace is ca 25-28m above modern lake level (Wang and Jiao, 1989). The sediments in these terraces must have been deposited during intervals during the Holocene and the late glacial when the lake was higher than today (Wang and Jiao, 1989).

The top of the T3 terrace has an elevation of ca 1117 m a.s.l. A 25-m high natural cutting through the terrace (T3 SW) ca 20m southwest of the lake consists of basal glacial meltwater deposits and fluvial sand deposits (5-25m), overlain by lacustrine clays (2-5m) and topped by beach sand and gravel (0-2m). Two samples from the lacustrine clay (3.0m and 4.3m) were radiocarbon-dated to 12,240±120 and 15,030±140 yr B.P. By extrapolating the sedimentation rate (0.0466 cm/yr) between these two dates, lacustrine deposition occurred between ca 17,600 and 10,100 yr B.P. Wang and Jiao (1989) estimate that the area of the lake corresponding to the 1117m shoreline would have been ca 330km^2.

Changes in water depth during the lacustrine phase are documented by changes in lithology. The basal part of the lacustrine unit (5.0-4.5m) in the T3 SW profile is blue-grey clay, characteristic of rather deep water. This unit dates to 17,600-15,460 yr B.P. The overlying sediment (4.5-3.8m) is black clay with abundant plant fragments. The increased organic content and the presence of plant fragments suggest decreased water depth between 15,460-13,960 yr B.P. The overlying sediment (3.8-3.0m) is blue-grey clay, suggesting an increase in water depth between 13,960-12,240 yr B.P. The overlying sediment (3.0-2.8m) is black silty clay. The presence of coarse detrital material suggests decreased water depth. This unit is dated to 12,240-11,810 yr B.P. The overlying sediment (2.8-2.4m) is blue-grey clay, suggesting increased water depth. This unit is dated to 11,810-10,950 yr B.P. The uppermost sediment (2.4-2.0m) is silty clay, suggesting decreased water depth. This unit is dated to 10,950-10,100 yr B.P.

Lake level dropped to below the level of the modern lake sometime around 10,100 yr B.P., thus creating the T3 terrace. Increased lake levels later in the Holocene resulted in lacustrine deposits being superimposed on the footslope of the T3 terrace.

These lacustrine deposits are represented in the T2 terrace. Dates of 8320±80 and 9650±130 yr B.P. on calcarous nodules from the surface of the T3 terrace in the T3 NE profile located to the northeast of the modern lake provide a further constraint on the age of the regression and subsequent transgression phases. These nodules are likely associated with soil formation on the T3 terrace surface, and therefore indicate that lake level must have dropped from 1117 m a.s.l. sometime before 9700 yr B.P. and never again rose to this elevation.

The top of the T2 terrace has an elevation between 1101-1105 m a.s.l. A profile (T2 NE-a) from the T2 terrace (1101 m a.s.l.) ca 1300m to the northeast of the modern lake consists of 1m of peat underlain by 0.65m of black lacustrine mud and 1.35m of grey lacustrine clay. The grey lacustrine clay indicates deeper-water conditions than the more organic, black lacustrine mud. Samples from 1.1m, 1.4m, 1.90m and 2.44m have been radiocarbon-dated to 6958±217, 8180±120, 9470±80 and 10,600±100 yr B.P. respectively. This suggests that the grey lacustrine clay was formed between 10,700-8820 yr B.P. and the organic black lacustrine mud between 8820-6760 yr B.P. There are four radiocarbon dates from the peat unit, showing that it was deposited between 1660 and 6760 yr B.P. The peat unit is of varying thickness and believed to have formed on the top of the terrace after lake level had fallen (Wang and Jiao, 1989). Thus, the T2 terrace provides evidence of lake levels more than 8m above the modern lake level between 10,700-6760 yr B.P. The dates on the lacustrine clays from the T2 NE-a profile suggest that the lake level drop that created the T3 terrace probably happened earlier than implied by extrapolation of the radiocarbon dates from the T3 profile itself and that lake levels were high again before the start of the Holocene.

There is a second profile (T2 NE-b) from the T2 terrace (1105 m a.s.l.) ca 1200m to the northeast of the lake. This basal deposits in this profile consists of a basal lacustrine clay, dated to 14,800±100 yr B.P. This unit appears to be equivalent to the lacustrine clays found in T3 and confirms that the T2 terrace is superimposed on older lacustrine materials. Younger lacustrine clays overlie the basal clay. These clays are undated but are believed to be equivalent to the clays formed during the early Holocene transgression and recorded in profile T2 NE-a. The surface of the T2 terrace is covered by peat up to 5m thick. A single sample from near the base of this peat has been radiocarbon-dated to 6690±200 yr B.P. The peat is thought to have formed after lake level dropped, eroding out the T2 terrace. Thus the date of the peat from the T2 NE-b profile confirms that lake level dropped, probably to levels below that of the modern lake, sometime around 6760 yr B.P.

A subsequent increase in lake level resulted in renewed lacustrine deposition on the footslope of the T2 terrace. These lacustrine sediments are exposed in the T1 terrace. The top of the T1 terrace has an elevation of 1094 m a.s.l. There are two profiles through the T1 terrace. A sample of grey lacustrine clay from the top of the T1 terrace in the profile ca 20m southwest of the lake was radiocarbon-dated to 2834±57 yr B.P. A second profile from the T1 terrace ca 700m to the northeast of the lake consists of 1m of peat underlain by grey lacustrine clay. A sample from the peat and two samples from the clay were radiocarbon-dated to 2281±130 yr B.P., 3033±118 and 3830±70 yr B.P. respectively. The peat unit varies in thickness and appears to have formed after lake level dropped below the elevation of the T1 terrace (Wang and Jiao, 1989). Thus, the evidence from the T1 terrace suggests that the lake was 1-2m higher than today between ca 4000 and ca 2800 yr B.P. and dropped sometime after 2800 yr B.P.

The data from the terraces can be summarised as follows:

Terrace	Site	Elevation of top (m a.s.l.)	^{14}C dates (yr B.P.)	Estimated lake level (m)
T1	T1 SW profile T1 NE profile T1 NE profile	1093-1094	2834±57, 3033±118, 3830±70	1 – 2
T2	T2 NE-a profile T2 NE-b profile	1101-1102 1105	6958±217, 8180±120, 9470±80, 10600±100 6690±200, 14,800±100	7 – 8 13
T3	T3 NE profile T3 SW profile	1117 1117-1120	8320±80 9650±130 12,240±120 15,030±140	25 25-28

There are two cores (CK1, CFK) from the northern part of the Chaiwopu Basin, ca 7km northwest of the modern lake. A 500m-long core (CK1), taken at an elevation of 1114.762 m a.s.l., provides sedimentary records back to the Pleistocene (Zhang and Wen, 1990). The alternation of lacustrine and fluvial-alluvial deposits in the CK1 core provides a record of long-term changes in lake status. Although we summarise the data from this core, we do not use this information to code lake status changes. A 14.6m-long core (Core CFK), taken 300m northeast of CK1, provides a record of the last 30,000 yr B.P. (Zhang and Wen, 1990). Reconstructions of changes in water depth are based on changes in lithology, ostracode assemblages (Huang, 1990) and the presence/absence of aquatics (Li and Yan, 1990) in this core. There are 2 Holocene radiocarbon dates from the core. One of these dates was made on a ca 80cm-thick sample and is therefore not considered to provide a reliable age estimate. The chronology is therefore based on 1 radiocarbon date (Wang and Jiao, 1989; Gu et al., 1990), palaeomagnetic measurements (Li et al., 1990) and the assumption that the top of the core can be dated by reference to the radiocarbon-dated age of the top of the CK1 core.

The sediments from CK1 are characterised by an alternation of lacustrine deposits (102.1-83.94m, 66.93-51.89m, 42.4-34.64m and 23.35-13.4m) and fluvial-alluvial units (550.68-102.1m, 83.94-66.93m, 51.89-42.4m, 34.64-23.35, 13.4-0m). The lacustrine units represent intervals when the lake was higher than today. The fluvial-alluvial units represent intervals when the lake was as low or lower than today. There are 8 U-series dates and 3 radiocarbon dates on the CK1 core. Three of the U-series dates and one of the radiocarbon dates are thought to be too young. The chronology of the core is therefore based on 5 U-series and 2 radiocarbon dates.

The first lacustrine unit (102.1-83.94m) is a laminated, grey-blue clay and silty clay, characteristic of deepwater deposition. The ostracode assemblage is initially (101.1-102.1m) dominated by *Candona* spp., *Ilyocypris biplicata*, *Limnocythere dubiosa* and *L. sancti-patricii*, species characteristic of fresh- to brackish-water conditions (Huang, 1990). In the middle part of the unit (93.0-93.20m) the assemblage is dominated by *Paracypricercus levis*, a fresh-water species (Huang, 1990). Two samples from 81-82m and 87-91m were U-series dated to 303,000±70,000 yr B.P. and 211,000±73,000 yr B.P. respectively. The date from 87-91m is clearly too young (Gu et al., 1990), but extrapolation from the date at 81-82m suggests this lacustrine phase occurred before ca 310,000 yr B.P.

The second lacustrine unit (66.93-51.89m) is yellow and blue clay and silty clay. No ostracodes were found in the unit. Two samples from 53.0m and 57.7m were U-series dated to 67,000±8000 yr B.P. and 211,000±90,000 yr B.P. The age of 67,000±8000 yr B.P. is probably too young (Gu et al., 1990). We estimate this unit was deposited between ca 251,000-194,000 yr B.P. by interpolation of the sedimentation rate (0.0215 cm/yr) between the date of 211,000±90,000 yr B.P. and a date of 62,000±7000 yr B.P. (23-24m) from an overlying unit.

The third lacustrine unit (34.64-42.4m) is grey clay and silty clay. The ostracode assemblage (39.6-39.85m) is dominated by *Paracypricercus levis*, *Candona neglecta, Candoniella albicans, C. mirabilis* and *Ilyocypris biplicata*, indicating fresh-water conditions (Huang, 1990). There are no dates from this unit, but interpolation between the available U-series dates on under- and over-lying units suggests it was formed between 147,000-110,500 yr B.P.

The fourth lacustrine unit (23.35-13.4m) shows considerable lithological variability. No ostracodes were found in the unit. Samples from 23-24m, 16.0m and 14.0m were U-series dated to 62,000±7000 yr B.P., 31,000±3000 yr B.P. and 20,000±2000 yr B.P. respectively, while a sample from 16.0m was radiocarbon-dated to 23,821±766 yr B.P. Given the discrepancy between the U-series and radiocarbon dates at 16.0m, we use only the U-series dates to erect a chronology for this unit. The unit can be divided into five sub-units. The first sub-unit (23.35-21.68m) is yellow lacustrine clay with fine sand lenses (ca 2cm thick), suggesting moderately deep water. This unit was deposited between 62,000-54,400 yr B.P. The second sub-unit (21.68-19.97m) is black fine sand with yellow clay, suggesting decreased water depth. This unit is dated to between 54,400-47,370 yr B.P. The third sub-unit (19.97-18.11m) is grey-blue laminated clay, suggesting increased water depth. This unit is dated to between 47,370-39,700 yr B.P. The fourth sub-unit (18.11-16.0m) is brown-yellow clay. The disappearance of laminations suggests decreased water depth. This unit can be dated to 39,700-31,000 yr B.P. The fifth sub-unit (16.0-13.4m) is grey-green laminated silty clay. The presence of laminations suggests increased water depth. This unit was deposited between 31,000-16,700 yr B.P.

The uppermost sediments in CK1 are fluvial-alluvial sand, gravel and stone (13.4-0.2m). A modern soil is developed in the upper 20cm of this unit. Two samples from 0.7-1.3m and 0.35m were radiocarbon-dated to 3696±90 and 3852±245 yr B.P. respectively, suggesting the fluvial-alluvial deposits were formed before ca 3700 yr B.P. The presence of a soil in the uppermost part of these sediments nevertheless shows that low lake levels persisted to the present day.

In summary, although the CK1 core cannot provide a high-resolution record of recent changes in lake level, it attests to multiple phases when the lake was higher than present during the Quaternary. Specifically, the lake was significantly higher than today before 310,000 yr B.P., between 251,000-194,000 yr B.P., between 147,000-110,500 yr B.P. and between 62,000-16,700 yr B.P. The sediments from the last lacustrine phase suggest that lake depth underwent significant changes through time. However, because of the limited number of available dates and the large uncertainties on the U-series dates, the chronology of these changes can only be estimated very approximately. We therefore make no attempt to correlate these changes with the record obtained from the CFK core.

The CKF core provides a record of the last ca 30,000 yr B.P. (Zhang and Wen, 1990). The basal unit (14.6-12.61m) is lacustrine clay. The unit can be sub-divided on the basis of changes in lithology and microstructure. The first subunit (14.6-14.2m) consists of grey silty clay, the second (14.2-12.61m) of blue-grey laminated clay, and the third

(12.61-12.21m) of pale-yellow silty clay. This sequence suggests an oscillation from shallower to deeper to shallower water conditions. It is not possible to determine the timing of these changes. The units are not radiocarbon-dated. Given that this unit is overlain by several meters of fluvial-alluvial sediments, extrapolation of the sedimentation rate between dates on overlying lacustrine units would not provide a plausible age estimate. Palaeomagnetic sampling of the core record (11 samples) suggests that the Lachamp polarity reversal (ca 30ka B.P.) can be identified near the base of the core (Li et al., 1990). Thus, the evidence suggest that lacustrine conditions occurred ca 30,000 yr B.P. and persisted for some unknown length of time thereafter.

The overlying sediments (12.21-3.6m) are coarse-fine sand, gravel and pebbles. According to Zhang and Wen (1990), these are fluvial-alluvial deposits. A sample from the upper boundary of this unit has been radiocarbon-dated to 9860±295 yr B.P. Thus, these sediments indicate that the lake was low before 9860 yr B.P.

The overlying sediments (3.6-3.0m) are black lacustrine sands and silts. The lithology is consistent with shallow-water or beach deposits. Given that the coring site is ca 7km from the modern lake, however, these deposits indicate that the lake was higher than today. We have assumed that the soil at the top of the CFK core began forming at the same time as the soil at the top of the CK1 core (i.e. ca 3700 yr B.P.). Interpolation of the sedimentation rate (0.058 cm/yr) between the radiocarbon-dated sample from the base of this unit and the core top, assumed to date to 3700 yr B.P., suggests that these beach deposits formed between 9860-8830 yr B.P. Since the elevation of the top of both the CK1 and CFK cores is ca 7m higher than the modern lake level, shallow water conditions at the core site imply that the lake was >7m higher than today. These deposits can therefore be correlated with the early Holocene lacustrine sediments in the T2 terrace, which imply water levels ca 7m higher than today.

The overlying sediment (3.0-1.1m) is black lacustrine clay. The change in lithology indicates increased water depth. The occurrence of *Typha* is consistent with a transition from beach to shallow-water deposition. This unit formed between 8830-5550 yr B.P.

The overlying sediment (1.1-0.35m) is pale-yellow lacustrine silty clay. The presence of silt-sized material suggests decreased water depth. This unit was formed between 5550-3700 yr B.P.

The sequence of changes shown by the lacustrine sediments in CFK (shallow-deeper-shallower) is the same as the sequence shown by lacustrine deposits in the T2 terrace. The timing of the changes is not the same. This may reflect the inadequacy of the chronology of the CFK core.

The soil that occurs at the top of the profile is assumed to correlate with the soil found at the top of CK1. Soil formation in both cores indicates that lake level was lower than the coring sites during the last ca 3700 yr B.P. This is consistent with the terrace evidence, which shows that the late Holocene increase in lake level (registered in T1) was only 1-2m.

In the status coding, very low (1) is indicated lake levels below that of the modern lake; low (2) by modern lake levels; moderately low (3) by lacustrine deposits in the T1 terrace (1-2m higher than modern lake level); intermediate (4) by organic muds in the T2 terrace (at ca 7-8m higher than modern lake level); moderately high (5) by clay in the T2 terrace (at ca 7-8m higher than modern lake level); high (6) by organic shallow-water deposits in the T3 terrace (ca 25m above modern lake level); and very high (7) by non-organic blue-grey clays in the T3 terrace (ca 25m above modern lake level).

Radiocarbon dates

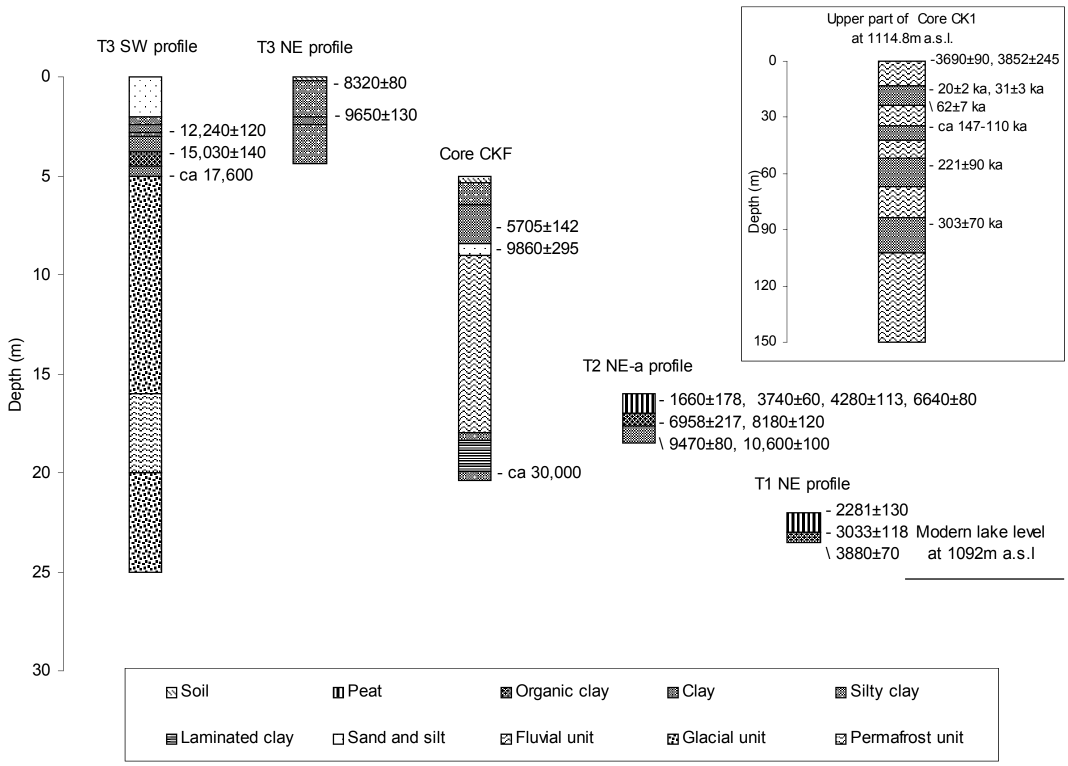
LB-127	23,821±766	16.0m, clay Core CK1 (not used, discrepancy with U-series dates)
LB	15,030±140	4.3m, clay, T3 terrace profile (SW)
LB	14,800±100	ca 5m, clay, T2 terrace profile (NE-b)
LB	12,240±120	3m, clay, T3 terrace profile (SW)
LB	10,600±100	2.44m, clay, T2 terrace profile (NE-a)
GC-87086	9860±295	3.6 m, sand ,Core CKF
LB	9650±130	ca 2m, calcareous nodules, T3 terrace profile (NE)
LB	9470±80	1.9m, clay, T2 terrace profile (NE-a)
LB	8320±80	ca 0.5m, calcareous nodules, T3 terrace profile (NE)
LB	8180±120	1.4m, mud, T2 terrace profile (NE-a)
LB	6958±217	1.1m, mud, T2 terrace profile (NE-a)
LB	6690±200	ca 3m, peat, T2 terrace profile (NE-b)
LB	6640±80	0.9m, peat, T2 terrace profile (NE-a)
LB-156	5705±142	2.85-3.6 m, clay, Core CKF (not used, sample size too large)
LB	4228±113	0.7m, peat, T2 terrace profile (NE-a)
LB-154	3852±245	0.7-1.3m, clay, Core CK1
LB	3830±70	ca 0.5m, clay, T1 profile (NE)
LB	3740±60	0.5m, peat, T2 terrace profile (NE-a)
LB-153	3696±90	0.35m, clay, Core CK1
LB	3033±118	ca 1.2m, clay, T1 terrace profile (NE)
LB	2834±59	ca 0.5m, clay, T1 terrace profile (SW)
LB	2281±130	ca 1.5m, clay, T1 terrace profile (NE)
LB	1600±178	0.3m, peat, T2 terrace profile (NE-a)

LB indicates the ^{14}C Lab of Lanzhou Institute of Glaciology and Geocryology, Chinese Academy of Science, and GC is the ^{14}C Lab of Geochemistry Institute, Chinese Academy of Science.

U-series dates

303,000±70,000	81-82m, carbonate, Core CK1
211,000±90,000	57.7m, carbonate, Core CK1
211,000±73,000	87-91m, carbonate, Core CK1 (possible ATY)
67,000±8000	53.0m, carbonate, Core CK1 (possible ATY)
62,000±7000	23-24m, carbonate, Core CK1
31,000±3000	16.0m, carbonate, Core CK1
20,000±2000	14.0m, carbonate, Core CK1
2500±?	1.25-1.35m, carbonate, Core CK1 (not used, discrepancy with radiocarbon dates)

The samples were dated by the Geochemistry Institute, Chinese Academy of Science.



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Coding

ca 30,000 yr B.P.	moderately high (5)
17,600-15,460 yr B.P.	very high (7)
15,460-13,960 yr B.P.	high (6)
13,960-12,240 yr B.P.	very high (7)
12,240-11,810 yr B.P.	high (6)
11,810-10,950 yr B.P.	very high (7)
10,950-10,100 yr B.P.	high (6)
??????-10,700 yr B.P.	very low (1)
10,700-8820 yr B.P.	moderately high (5)
8820-6760 yr B.P.	intermediate (4)
6760-4000 yr B.P.	very low (1)
4000-2800 yr B.P.	moderately low (3)
2800-100 yr B.P.	very low (1)
2800-0 yr B.P.	low (2)

Preliminary coding: 21-01-1999
Second coding: 01-02-1999
Final coding: 24-01-2001
Coded by GY and SPH

3.27. Lop Basin, Xinjiang Autonomous Region

Lop Basin (also known as Lop Nur: 39°45’-40°50’N, 90°10’-91°25’E, 780-795 m a.s.l.) lies in the eastern part of the Talim Basin and forms the lowest part of that basin. The catchment area for the Lop Basin is ca 300,000 km^2. The Tarim Basin is of tectonic origin and was formed during the Neogene/Quaternary. The basin bedrock is mainly sandstone. The Lop Basin has been the sink of the Tarim Basin since the Neogene. Drilling surveys indicate that there are several hundred meters of lacustrine sediments in the Lop Basin and suggest that a lake was present in the Lop Basin throughout the Quaternary (Yan, et al. 1998). There is no lake today and Lop is now a playa. Before the Tang Dynasty (618-907 A.D.), however, the Lop consisted of three lakes: Taitema Lake (88 km^2, 807 m a.s.l. of the lake floor), Kalaheshun Lake (1100 km^2, 788 m a.s.l.) and Lop Lake (5350 km^2, 778 m a.s.l.). Lop Lake was the lowest lake of the three lakes, and the other lakes were connected to it by natural channels. Lop Lake was fed by the Peacock River, Taitema Lake was fed by the Talimu River, and Kalaheshun Lake was fed by the Churchen River. These rivers drained the Tarim basin. The volume of these rivers decreased after the Tang Dynasty period because of intensive use for irrigation coupled with increasing aridification of the region. The lakes became completely dry during this century (Yan and Shao, 1993). Although there are glaciers in the mountains surrounding the Lop Basin, glacial meltwater does not contribute significantly to the water budget of the basin. The annual precipitation in the Lop Basin is 20 mm and potential evaporation is 2800-3000 mm. Given the large catchment area, it is possible to maintain a lake area of ca 5000 km^2 in the shallow Lop Basin with this amount of precipitation (Yan and Shao, 1993). There are extensive aeolian deposits within the basin and the vegetation cover is sparse.

There are three cores from the Lop Basin (K1, L4 and F4). The K1 core is 100.2 m long and was taken from the centre of the Lop playa. It provides a lithological and pollen record back to the early Pleistocene. The lowest part (100.2-66.2 m) of the K1 core, covering the Early Pleistocene, consists of greenish-grey to brown-grey mudstone; the middle part (66.2-17.4 m), covering the Middle Pleistocene, is characterized by gypseous greenish-grey to greyish-green mudstone initially and greyish-green mudstone with gypsum interbeds subsequently. This lithology suggests that the Middle Pleistocene was relatively dry in the Lop region (Yan, et al. 1998). The upper part (17.4-0 m) of the K1 core consists of brown and grey clay, indicating wetter conditions during the Late Pleistocene. There are 3 radiocarbon dates from the top 6m of the K1 core, indicating that this part of the record covers the last ca 26,000 years. Although the sediments in this part of the core show significant changes in lithology from clays to sands to salt-rich layers (Yan et al., 1998), which presumably reflect changes in lake depth, the information given in the published paper is insufficient to construct a detailed record of these changes.

The remaining two cores (L4: Yan et al., 1983; F4: Zhen et al., 1991; Wu, 1994), taken ca 20 km apart in the central part of the Lop Basin, provide a lithological record back to ca 20,800 and 19,670 yr B.P. respectively. Core L4 is 8.83 m long and was taken at the southern end of the Lop playa. Core F4 is 4.5 m long and was taken to the east of the Lop playa. There are 4 radiocarbon dates for F4 and 3 dates for L4. The stratigraphy in both cores is complex and correlations between the cores are not always straightforward. This partly reflects the fact that the records were described by different authors and with different levels of detail, but may also be due to geomorphic changes in the low-lying

Lop Basin which resulted in changes in the shape of the lake basin through time. Nevertheless, we have reconstructed a composite record from the two cores which seems to reflect climatically-driven change in water depth through time. The changes in water depth are based on changes in lithology and aquatic pollen assemblages from both cores and the ostracode record from F4. The reconstructed changes are broadly consistent with the interpretation of the original authors. The chronology is based on the 4 radiocarbon dates from core F4 and the 3 dates on core L4.

The basal unit (8.83-8.5 m), only represented in L4, is black mud. The lithology indicates lacustrine conditions in the basin. The organic nature of the sediments, indicated by the black colour, suggests the lake was not particularly deep. The presence of *Typha* is consistent with shallow lacustrine conditions. A sample from near the base (8.83-8.5 m) of this unit is radiocarbon dated to 20,780±300 yr B.P. By linear interpolation between this date and the date at 3.1 m in L4, this phase occurred ca 20,800-20,470 yr B.P.

The overlying unit (8.5-8.1 m), again only represented in L4, is black muddy silt. The change in lithology suggests the lake became somewhat shallower. There is a single grain of *Typha* in the pollen spectrum. By interpolation, this phase occurred ca 20,470-19,650 yr B.P.

The overlying units (8.1-4.1 m in L4; 4.5-2.13m in F4) consist of clayey silts and silty clays, containing varying amounts of carbonate and gypsum. Three units are recognised in each core. In L4 the sequence is greyish yellow to greyish black silty clay (8.1-5.2m: 19,650-13,680 yr B.P.), overlain by greyish black silt (5.2-4.8 m: 13,680-12,860 yr B.P.) and finally greyish yellow silt (4.8-4.1 m: 12,860-11,420 yr B.P.). This sequence appears to represent an initial increase in water depth followed by gradual shallowing of the lake. The sequence from F4 also appears to represent a gradual shallowing. Thus, the basal unit (4.5-3.75 m: 19,670-17,000 yr B.P.) in F4 is discontinuously laminated, varicoloured (greyish red- brown yellow) clay and silty clay containing carbonate with some crumb-like gypsum. The presence of the halophyte *Cyprideis littoralis* (Brady) is consistent with the idea that the lake was saline. This unit (3.75-2.75 m: 17,000-13,290 yr B.P.) is overlain by brown clayey silt with scattered gypsum. The uppermost unit (2.75-2.13 m: 13,290-11,020 yr B.P.) is gypseous, brown silty clay. Thus, although the lithological description of the individual units is not identical, possibly because of the use of different terminologies by the authors of the original papers, it would appear that the sequence of changes in water depth indicated by both the F4 and L4 records are the same. The records show that this shallowing began sometime between 19,650 yr B.P. and culminated between 11,420 (L4) and 11,020 (F4) yr B.P.

The overlying units in both cores represent the culmination of this shallowing. In F4, the overlying unit (2.13-1.87 m: 11,020-10,080 yr B.P.) is described as a greyish-green clay, with scattered tablet gypsum, high-Mg calcite, dolomite and occasional scattered halite. The mineralogy indicates that the water salinity increased markedly and suggests the lake was very low. The presence of halophytic ostracodes (including *Cyprideis littoralis* (Brady), *Cyprinotus* sp. and *Eucypsis inflata* (Sars)), is consistent with highly saline, shallow-water conditions. The overlying unit (4.1-3.1 m) in L4 is greyish-black clayey silt with thin sand interbeds. The presence of sand indicates the lake became shallower. A sample from the top of the unit (3.1 m) is radiocarbon dated to 9360±120 yr B.P., suggesting this shallow phase occurred 12,440-9360 yr B.P. Thus, although the lithology of the overlying unit in each core is apparently different, the sequence and approximate timing of the change in recorded water depth is apparently the same.

The overlying unit (1.87-0.78m: 10,080-6390 yr B.P.) in F4 is a fine-grained (clay or mud) gypseous deposit. The change in lithology and mineralogy indicates an increase in water depth, as does the absence of halophytic ostracodes. Wu (1994) divided this unit into five sub-units, with clay alternating with sediments described as mud. Each unit is described as dark, black and greyish-black to greyish-brown in colour, contains only tablet gypsum, and is characterised by the absence of halophytic ostracodes. Given that there is no change in colour, mineralogy or ostracodes, it seems unlikely that there was a significant change in water depth. Furthermore, the whole unit can be correlated with the unit in L4 (3.1 -2.7 m:9360-7920 yr B.P.) described as greyish-black mud.

The overlying unit (2.7-2.4 m) in L4 is a light-yellow, fine sand. The change in lithology indicates a marked shallowing. Interpolation between the radiocarbon dates from L4 suggests this shallowing phase occurred 7920-6840 yr B.P. The unit is not recorded in F4, and may therefore reflect localised geomorphic changes in the lake basin.

The overlying unit (0.78 m-0.23 in F4: ca 6390-3060 yr B.P.; 2.4-1.7 m in L4: ca 6840-3970 yr B.P.) is silty clay/clayey silt. The sediments in F4 are recorded as containing carbonate and gypsum. Assuming that the sand layer, which is only recorded in L4, does not represent a climatically-induced change in water depth, the transition from muds/clays to silty clays indicates decreased water depth.

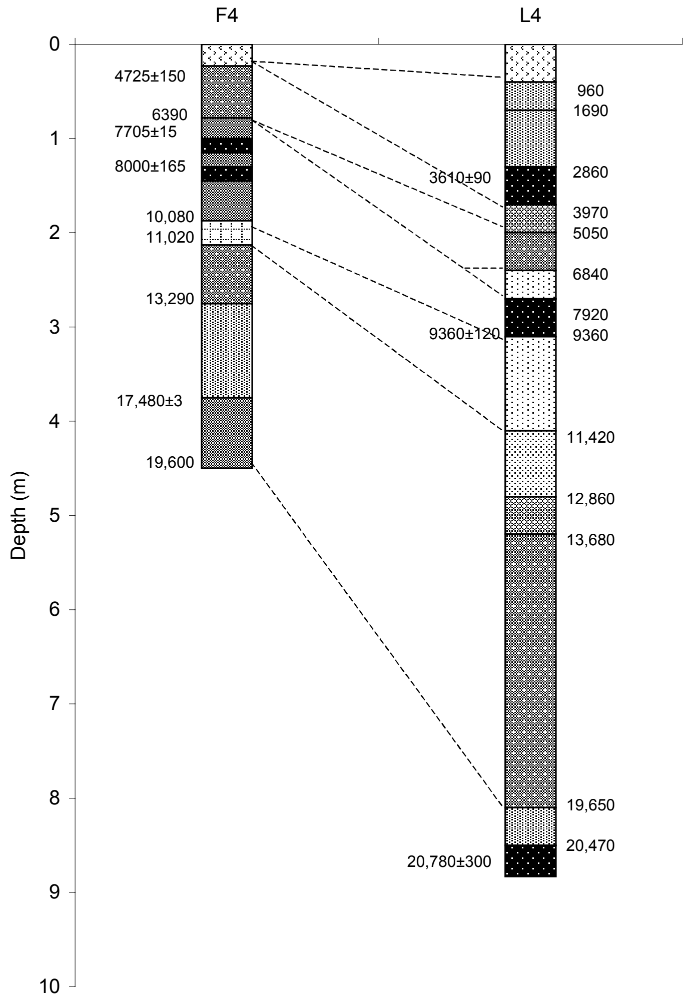
The overlying unit (1.7-1.3 m), which is only recorded in L4, is black mud, containing granulate-like gypsum. The transition to black mud suggests the lake became deeper. A sample from 1.5 m is radiocarbon dated to 3610±90 yr B.P. By extrapolation, the upper boundary of the mud unit is dated to 2860 yr B.P.

The overlying unit (1.3-0.7 m), which is only recorded in L4, is black muddy silt. The change in lithology suggests shallowing. Assuming the top sediments are modern, this phase occurred ca 2860-1690 yr B.P.

The overlying unit (0.7-0.4 m), which is only recorded in L4, is yellow clayey silt. The change in lithology and organic content (as reflected by the colour) suggest water depth increased. This phase occurred ca 1690-960 yr B.P.

The uppermost unit in both cores (0.23 m-0 in F4; 0.4 m-0 in L4) is halite. The presence of this salt crust indicates that the lake became a shallow playa. This dramatic shallowing likely explains the absence from F4 of the sediments deposited between ca 4000 and 960 yr B.P. recorded in L4. On the basis of the dates on core L4, the halite crust started forming ca 960 yr B.P. This is consistent with the historical records from Lop Basin.

In the status coding, very low (1) is indicated by halite deposits; low (2) by gypsum, dolomite and halite with halophytic ostracodes in F4 and sand layers in core L4; intermediate (3) by muddy silts in both cores; high (4) by silty clays or clayey silts in both cores; and very high (5) by clays and/or muds in both cores.



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Radiocarbon dates

26,172±479	ca 6.0m, clay, K1 (not used)
23,668±347	5.5 m, clay, K1 (not used)
20,780±300	8.83-8.5 m, mud, L4
17,480±300	3.9 m, clay, F4
9360±120	3.1 m, mud, L4
9220±174	1.15 m, clay, K1 (not used)
8000+165/-160	1.3 m, mud, F4
7705±150	1.0 m, mud, F4
4725+150/-145	0.5 m, clay, F4
3610±90	1.5 m, mud, L4

Coding

20,800-20,470 yr B.P.	very high (5)
20,470-19,650 yr B.P.	intermediate (3)
19,670-11,020 yr B.P.	high (4)
11,420-9360 yr B.P.	low (2)
10,080-6390 yr B.P.	very high (5)
6840-3060 yr B.P.	high (4)
3970-2860 yr B.P.	very high (5)
2860-1690 yr B.P.	intermediate (3)
1690-960 yr B.P.	high (4)
960-0 yr B.P.	very low (1)

Preliminary coding: February 1996
Third coding: January 1999
Final coding: 24-01-2001
Coded by BX, GY and SPH

3.28. Manasi Lake, Xinjiang Autonomous Region

Manasi Lake (also called Manas Lake; 45.45°N, 86.0°E, 251m above sea level) is a now-dry salt lake in a closed basin in the Zhungeer (Zhunggar) Basin, western Xinjing. The Zhungeer Basin is bounded to the north by the Altai Shan Mountains, to the south by the Tian Shan Mountains, and to the west by a series of mountains and plateau that rise to elevations of ca 3000m. The bedrock geology in these mountain ranges consists of granites, Devonian and Carboniferous sedimentary rocks, and Mesozoic limestones. The Zhungeer Basin is covered by Quaternary sedimentary deposits. The Manasi Basin is inset within the central part of the Zhungeer Basin, and originated as a fault depression (Lin et al., 1996). The lake was originally fed by the Manasi River, which flows from the northern slopes of the Tian Shan Mountains (a distance of ca 450 km) and entered the lake basin from the southwest (Lin et al., 1996). Thus, the lake had a large catchment area (ca 11,000 km^2), stretching from the Tian Shan and covering much of the central part of the Zhungeer Basin. The Manasi River is largely fed by the seasonal melting of glaciers in the upper parts of the Tian Shan, and by local precipitation. Local precipitation within the Manasi basin itself provided some additional input to the lake, but there is no evidence that groundwater inflows contributed to the lake water budget (Huang et al., 1987). Geomorphic data suggest that Manasi Lake may have been fed by rivers draining the Altai Shan at an earlier stage in its history (XCET, 1978; Rhodes et al., 1996). According to Rhodes et al. (1996), there may be some seepage of water from a small, freshwater closed-basin lake (Lake Ailike) which is still supplied by runoff from further north. The increasing use of the water from the Manasi River for irrigation during the late 1950's resulted in the total diversion of the major source of inputs to Manasi Lake, and the lake has been dry since about 1960 (Huang et al., 1987). In 1957, the lake had a water depth of ca 6m (257 m a.s.l.) and an area of ca 750 km^2 (Lin et al., 1996). The lake water was hypersaline (Wei and Gasse, 1999). The former lake bed is now covered by a halite salt crust.

The Zhungeer desert region lies within the zone of influence of the Westerlies, and receives rainfall during the autumn and winter months. It lies well beyond the present western limits of the Pacific monsoon, and beyond the northern limits of the Indian monsoon. Precipitation in the Zhungeer Basin is generally low: in the central part of the basin, in the vicinity of Manasi Lake, the mean annual precipitation is ca 100 mm (Lin et al., 1996) but somewhat higher precipitation levels are registered towards the margins of the basin. At Urumqi in the foothills of the Tian Shan, for example, the mean annual precipitation is ca 276 mm (Watts, 1969). Mean annual evaporation is ca 10 times larger than precipitation (ca 1000 mm at Manasi, ca 2100 mm at Urumqui; Huang et al., 1987). The mean annual temperature is ca 4°C, but with a large seasonal amplitude from ca -16°C in January to ca 24°C in July. The vegetation in the Manasi Lake region is typical of desert climates, and is dominated by Chenopodiaceae (including *Salsola, Halocnemum, Halostachys* and *Kalidium*) and Polygonaceae (including *Calligonum*) (Huang et al., 1987).

Manasi Lake has been considerably more extensive in the past. Carbonate deposits from an exposed lacustrine sequence ca 20km southwest of the modern salt flat have been radiocarbon-dated to 5310$\pm$95 yr B.P., indicating the lake was more extensive in the middle Holocene (Rhodes et al., 1996). Interpretation of satellite imagery suggests that Manasi Lake may have been connected formerly to Aibi Lake (Rhodes et al., 1996), which lies ca 180-200 km to the west-southwest. However, this connection has not yet been verified in the field. Although shoreline features dated to the early-to-mid Holocene occur around Aibi Lake, there are no apparent relict shorelines around Lake Manasi. However, the Late Quaternary history of Manasi Lake has been reconstructed from three 5-more long cores from the centre of the now-dry lake (Sun et al., 1994; Lin et al., 1996; Rhodes et al., 1996; Wei and Gasse, 1999).

The three 5m-long cores (LM-I, LM-II and LM-III) were all take a very short distance (< 1m) apart in the centre of the lake (Lin et al., 1996). The stratigraphy of the three cores is very similar and they can therefore be easily cross-correlated (Lin et al., 1996). Core LM-II has been studied for lithology, mineralogy, organic content and geochemistry, diatom assemblages and for the stable isotopes of oxygen and carbon (Lin et al., 1996; Rhodes et al., 1996; Wei and Gasse, 1999). Past environmental changes in the Manasi Basin, including changes in the size of the lake, have been interpreted on the basis of these indicators (Rhodes et al., 1996). Pollen analysis has been carried out on Core LM-I and used to reconstruct the palaeovegetation and palaeoclimatic history of the basin (Sun et al., 1994).

There are eight AMS-radiocarbon dates on bulk sediments from Core LM-II, and a single conventional date from both core LM-I and from core LM-III (Sun et al., 1994; Lin et al., 1996; Rhodes et al., 1996). The near-basal AMS date (4.66-4.68m) on carbonate material in Core LM-II is 32,100$\pm$750 yr B.P. This age is close to the reliability limits for radiocarbon dating. Sun et al. (1994) do not use this near-basal date to erect their chronology, preferring to extrapolate the sedimentation rates from the overlying (Holocene) ages to date the base of the core. As a result, they assume that the core covers only the last 14,000 years. Lin et al. (1996) and Rhodes et al. (1996) point to the fact that the date of 32,100 yr B.P. is consistent with an age of 37,800$\pm$1500 yr B.P. from the same unit at a depth of 4.684m in Core ML-III. Thus, while cautioning that this date may be too old, they suggest that the sedimentary record from Manasi Lake covers the last ca 38,000 years.

Here we reconstruct the changes in water depth and salinity on the basis of a consensus interpretation of the changes in lithology, mineralogy, organic content, geochemistry, and aquatic pollen assemblages. We do not use changes in diatom assemblages as an indicator of changes in water depth and/or salinity, since the sporadic presence of diatoms appears to reflect changes in water chemistry consequent on discrete flushes of halite from the catchment. Most of the lake bottom deposits are calcareous, but there are changes in the relative abundance of calcite and aragonite forms of the calcium carbonate. We interpret the dominance of calcite in the lake deposits to indicate relatively freshwater conditions and the dominance of aragonite to indicate more saline conditions (Lin et al., 1996). Changes in the stable carbon isotope composition can be interpreted in a similar fashion, since shifts in $\delta^{13}C$ ($^{13}C/^{12}C$) reflect the evaporative concentration (and hence water salinity) in a closed basin under arid climate conditions (see e.g. Zhang, 1985; Gasse et al, 1987; Wei and Gasse, 1999). Higher $\delta^{13}C$ values indicate more saline conditions while lower $^{13}\delta C$ values indicate fresher water conditions. The abundance of amorphous (AOM) and lignocellulosic (LOM) organic matter, as identified by light microscopy, indicates the relative contribution of algae (AOM) and higher (LOM) plants (see e.g. Talbot and Livingstone, 1989). Changes in the AOM/LOM ratio provide an indirect indication of water depth, with algal productivity and hence abundant AOM indicating relatively deepwater conditions. Our chronology is based on the assumption that the near-basal dates of 32,100 and 37,800 yr B.P. are reliable. The two dates from ca 4m within the core show a dating reversal and overlapping ages at one standard deviation; we use the date of 10,120±100 yr B.P. (3.979-3.999m) to erect a chronology because this date has smaller error bars.

There are some discrepancies in the descriptions of the stratigraphy and lithology given in the original papers. Specifically, the stratigraphic log given in Rhodes et al. (1996) shows a stratigraphic gap, presumably due to non-recovery of core material rather than a sedimentary hiatus, between 4.76-4.90m. Two sections of the core (between 3.12-3.24m and between 2.14-2.33m) are described in the log as "polluted", presumably indicating that the sediments were contaminated or perturbed during the coring operation. However, Lin et al. (1996) do not mention these gaps or polluted zones and give stratigraphic descriptions covering these intervals. Furthermore, the analytical results given in Sun et al. (1994), Lin et al. (1996) and Rhodes et al. (1996) appear to include samples from these intervals. We therefore assume that the disturbance to the core material was minimal, and did not substantially affect the record. A further difficulty is presented by the fact that there are differences (of up to 50 cm) in the depths quoted for individual stratigraphic boundaries between different papers, and smaller (ca 5-10 cm) differences even within the same paper. We have assumed that the stratigraphic log given by Rhodes et al. (1996) is the most detailed record of the depths of individual units, and have supplemented this information with data from Lin et al. (1996) wherever possible.

The basal sediments (4.60-4.98m in Core LM-II, below 4.70m in Core LM-I) are silty clays of lacustrine origin. The abundance of fine detrital material (60-80%) is consistent with moderately deep water conditions. Evaporitic minerals (e.g. chloride) occur only at very low abundances (<3%), again consistent with fresh water conditions. The carbonate component of the sediments is predominantly calcite, consistent with relatively fresh and deepwater conditions. The carbonate content increases upcore and reaches a maximum (ca 28%) at 4.67m. The relative abundance of amorphous organic matter (AOM) in this basal unit suggests that the carbonate deposition was enhanced by algal photosynthetic activity. Again, the presence of aquatic algal communities supports the idea that the lake was relatively fresh and deep. However, the abundance of lignocellulosic (LOM) organic matter increases towards the top of the unit, indicating increased inputs from the catchment and possibly a trend towards shallower conditions. The relatively low $\delta^{13}C$ values (-6 to -5‰) are also consistent with fresh and relatively deepwater conditions. The aquatic pollen assemblage is characterised by *Typha* and *Sparganium*. A sample from this unit in Core LM-II (4.664-4.684m) was radiocarbon-dated to 32,100±750 and a sample from 4.684m in Core LM-III was radiocarbon-dated to 37,800±1500 yr B.P. Assuming these dates are reliable, the evidence suggest that the lake was relatively high before ca 32,000 yr B.P.

The overlying unit (4.18-4.60m. in Core LM-II) is red-brown, mottled clay. The sediments consist primarily of detrital minerals (59-87%) and the carbonate content is relatively low (<10%). Between 4.60-4.40m, there is an increase in the amount of extremely fine grain-sized (6-14µm) material. Lin et al. (1996) and Rhodes et al. (1996) interpret this fine material to reflect aeolian inputs to the sediments. Bands of coarse quartz and feldspar sand at 4.30m and again at 4.22m are considered to indicate increased fluvial inputs to the lake, possibly as a result of flash floods (Rhodes et al., 1996). The presence of the fine-grained aeolian dust and of the coarse fluvial material do not seem to be diagnostic of changes in water depth. However, changes in the nature of the primary matrix material are diagnostic of a change in water depth. Specifically, the colour and the mottling indicate oxidation of the sediments, and are consistent with relatively shallow water or possibly even dry conditions. The organic fraction within the unit consists primarily of altered, reworked higher plant debris (LOM) and there is little amorphous, algal organic matter (AOM). Again, this is consistent with shallowing. The increase in $\delta^{13}C$ value (-3‰) suggests increased salinity. Terrestrial pollen concentrations are low (<2000 grains/cm^3) and there is no aquatic pollen present, as might be expected if the lake became shallow and dried out. The deposit is indurated, consistent with at least intermittent desiccation of the lake. The presence of halite and of significant amounts of gypsum also supports the idea of shallow conditions. The extremely low sedimentation rate (0.00312 cm/yr) between the near-basal date of 32,100 yr B.P. (4.664-4.684m) and the date of 10,120 yr B.P. on the overlying unit (3.979-3.999m) is also consistent with dry conditions and the existence of sedimentary hiatuses. The evidence supports the idea that Lake Manasi was a playa lake, and at least intermittantly dry. It is difficult to date this interval of playa conditions. The oxidised and indurated nature of the sediments, and the presence of fluvial sands, suggest that deposition was intermittent. It is highly likely that material was removed by aeolian activity when the lake was dry. Extrapolation of the sedimentation rate between dates on the overlying units (0.00132 cm/yr) suggests that deposition of the overlying unit may only have begun ca 11,600 yr B.P. The dry phase could thus have persisted from ca 32,000 to 11,600 yr B.P., but it could also have been considerably shorter. Rhodes et al. (1996) tentatively attribute this dry phase to the Last Glacial Maximum, but it could equally well have been confined to the Late Glacial. The chronology of this playa phase will remain uncertain until additional radiocarbon dates can be obtained.

A return to lacustrine conditions is indicated by deposition of brown and grey carbonate clays (4.03-4.18 m in Core ML-II). The carbonate content peaks at ca 17% near the base of this unit and evaporite minerals are in low abundance (2-3%), consistent with fresh water conditions. The carbonate fraction is dominated by calcite, consistent with relatively fresh conditions, but aragonite is present in low amounts. Amorphous organic matter increased in abundance, consistent with a return to lacustrine conditions. The $\delta^{13}C$ values decrease (-3 to -8‰), indicating freshening and increased water depths. The terrestrial pollen concentration increased compared to the underlying unit, although it is still relatively low, but this increase and the presence of aquatic pollen (*Typha* and *Sparganium*) is consistent with increased water depths. According to the stratigraphic log in Rhodes et al. (1996), these deposits are laminated, which is consistent with relatively deepwater. However, Rhodes et al. (1996) suggest that the laminations are somewhat indistinct and the sediments show colour mottling. Thus, the interpretation of the water depth during this interval relative to the water depth during the initial lacustrine phase is not straightforward: the occurrence of laminations suggest that the water depth was greater than formerly but the presence of colour mottling suggests that the water was not as deep as during the initial lacustrine phase. The presence of aragonite here (and its virtual absence from the basal lacustrine unit) suggest that the lake was less deep than formerly. Since the colour mottling could have resulted from post-depositional processes, and the amount of aragonite present is small, we have interpreted the unit as indicating lacustrine conditions similar to those characteristic of the initial lacustrine phase. A sample from the transition between this unit and the overlying unit (3.979-3.999 m) is radiocarbon-dated to 10,120±100 yr B.P., suggesting this unit was deposited before ca 10,350 yr B.P.

The overlying unit (4.03-3.98) consists of fine sand beds interbedded with dark mud. These sands are interpreted as indicating significant fluvial discharges into the lake. The occurrence of significant freshwater pulses is consistent with the very low $\delta^{18}O$ values (- 10.45‰), which indicate that evaporative concentration was negligible. This unit is assumed to reflect conditions during a short interval, ca 10,350 to 10,100 yr B.P., when the lake was low.

The overlying sediment (3.98-3.40m) is soft, lacustrine carbonate clay (marl). The carbonate content increases progressively through the unit, and the stable isotope records broadly parallel the carbonate record. The unit can be subdivided into three phases: an early phase characterised by increasing water depth, a middle phase when the lake reached maximum water depths, and a third phase when water depth was decreasing. During the first phase (3.98-3.86 m), the sediments consist of non-laminated or poorly-laminated, olive grey, clay marl. The carbonate content of the sediments increased (from 5 to 20%). Calcite was the dominant form. The increasing carbonate content and the dominance of calcite are both consistent with increasing water depth. Aquatic pollen (*Typha* and *Sparganium*) is present at low abundances (< 2%) in the basal part of this unit, and subsequently virtually disappears. This is consistent with a gradual increase in water depth. During the second phase (3.86-3.70 m), the sediments consist of dark-coloured, finely-laminated clay marls. The preservation of laminations is consistent with increased water depth. The carbonate content continued to increase, and calcite remained the dominant component. Thus, the carbonate mineralogy is consistent with increased water depths. The sediments are highly organic, with both well-preserved amorphous and lignocellulosic components, consistent with relatively deepwater conditions. Decreased water depth during the third phase (3.70-3.40 m) is indicated by deposition of poorly-laminated, olive grey, clay marls. The carbonate content increased further, but aragonite was the dominant component. The shift in the relative importance of calcite and aragonite is consistent with increased evaporative concentration and, hence, decreased water depths. Increased levels of opaque lignocellulosic organic matter are consistent with increased inputs of weathered material from the catchment, and therefore decreased water depths. Rhodes et al. (1996) suggest that the association of organic matter with pyrite indicates deposition in stagnant, poorly-oxygenated water. This could also be consistent with shallower conditions. On the basis of a radiocarbon date of 10,120±100 yr B.P. from the lower boundary of these units at a depth of 3.979-3.999m (H-689) and a date of 7210±100 yr B.P. from a depth of 3.596-3.616m (H-605), the first lacustrine phase can be dated to between ca 10,120 and 9160 yr B.P. The second phase of deeper water can be dated to between 9160 and 7930 yr B.P. The final lacustrine phase is dated to between 7930 and 5630 yr B.P.

The overlying unit (3.40-3.15 m in Core ML-II) consists of sandy clay and coarse sands, interrupted by some centimeter-thick dark clay bands. This unit has been interpreted as indicating fluvial deposition, and marks a significant lake regression. The sands are predominantly quartz and feldspar, derived by long-distance transport from the mountains. The clay assemblages are dominated by detritic illite and halite, consistent with deposition of weathered material from the catchment. Thus, the mineralogy of the sediments is consistent with a fluvial origin. The abrupt decrease in ^{18}O values is consistent with high volume of freshwater inputs, and is also considered to be consistent with flushes of fluvial freshwater. The absence of aquatic pollen is consistent with fluvial deposition, as is the lack of evaporitic minerals and organic matter. The occurrence of gypsum in the lowermost part of this unit (ca 3.38m) is consistent with evaporative concentration as the lake shrunk. A single radiocarbon date of 4500±80 yr B.P. (H-690) has been obtained from a depth of 3.244-3.264m within this unit. Interpolation of the sedimentation rates between this date and over/underlying radiocarbon dates indicates that this phase of fluvial deposition occurred between 5630 and 3700 yr B.P. Fluvial deposits can be deposited relatively rapidly and extrapolation of the sedimentation rate from the overlying units suggests that cessation of fluvial deposition occurred as early as ca 4250 yr B.P.

The overlying unit (3.15-2.40m in Core ML-II) consists of an alternation between laminated carbonate clays, non-laminated carbonate clays and fluvial sands. The laminated clays indicate relatively deepwater conditions, the non-laminated clays indicate shallower conditions, and the fluvial sands indicate very shallow conditions. The sand layers (which are not shown on the sedimentary log) appear to occur at ca 3.10m, 2.90-2.80m and 2.50-2.40m. According to the sedimentary log, dark-coloured, finely laminated clays occur between 3.12-2.96m and again between 2.74-2.80m. Discrepancies between the published descriptions and the stratigraphic log make it difficult to correlate the changes in e.g. mineralogy, organic composition and stable isotopes precisely with the stratigraphic and lithologic changes. Nevertheless, there is a general increase in carbonate content through the lower part of this unit, to ca 2.90m, consistent with the prevalence of laminated lacustrine sediments. The maximum abundance of detrital organics and detrital minerals occurs at ca 2.81 m, in association with the major fluvial sand unit (2.90-2.80 m). The carbonate content reaches a maximum of between 30-40% in the lower part of the unit. The carbonate content decreases above ca 2.80m, consistent with the prevalence of non-laminated lacustrine sediments in the upper part of this unit. Finally, the highest $\delta^{18}O$ values occur in the non-laminated clays between 2.70-2.50m. Broadly speaking, this unit can be interpreted as indicating an initial phase of relatively deepwater conditions, interrupted by a single short-lived fluvial episode. A second phase of lacustrine conditions is both initiated and terminated by fluvial deposition. A sample from 2.82-2.84m is radiocarbon-dated to 3440±120 yr B.P., suggesting this unit was deposited between 4250-2580 yr B.P. Assuming that the first fluvial episode, marked by sands at ca 3.10m, represents a negligible interval of time, the first phase of deepwater conditions (indicated by laminated sediments) occurred from 4250 to 3460 yr B.P. The first major fluvial phase, which indicates a significant lake regression, occurs ca 3460-3380 yr B.P. A return to deepwater conditions, marked by laminated clays, occurred between 3380-3260 yr B.P. The deposition of non-laminated clays indicates shallower lacustrine conditions between 3260-2780 yr B.P. The final major fluvial phase occurred ca 2780-2580 yr B.P.

The transition to the overlying unit is marked by a sharp peak in halite (2.40m). Rhodes et al. (1996) suggest this halite was derived from dissolution of salt crusts around the margin of the lake as the lake rose during the return to lacustrine conditions that followed the end of the fluvial phase.

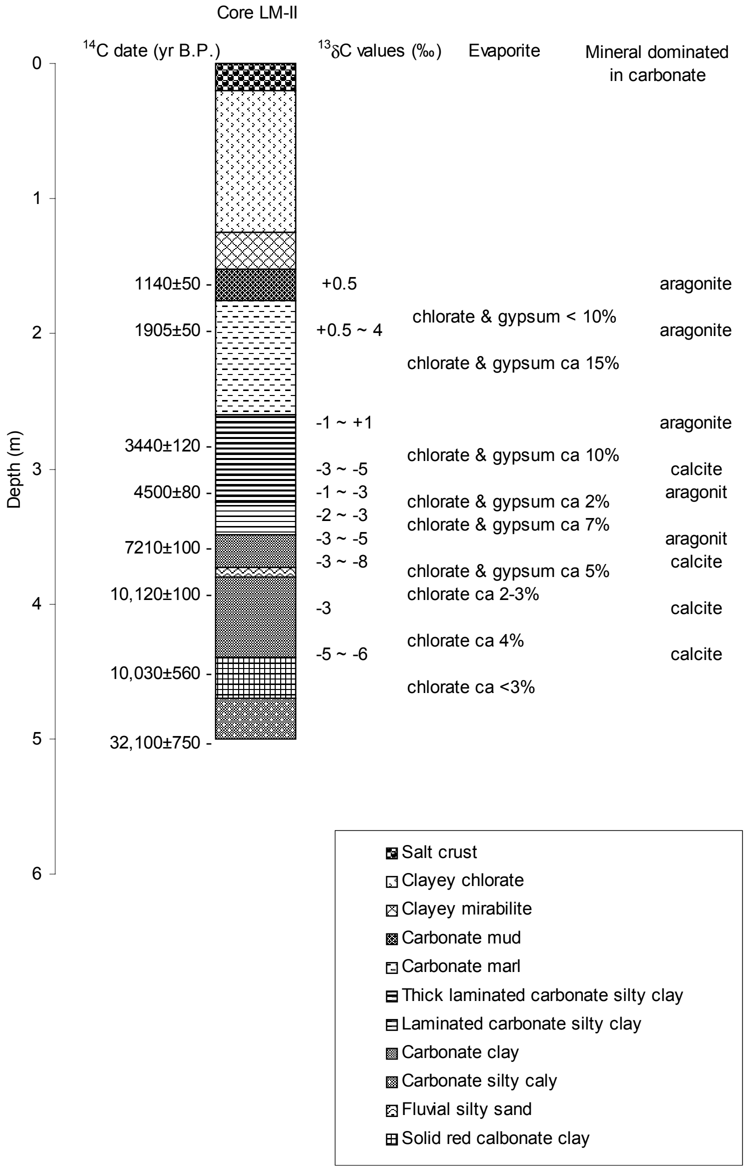
The overlying unit (2.40-2.05m) is dark-coloured, sapropelic mud. The carbonate content fell to <15% and calcite was the dominant form. The lithology indicates relatively deepwater. The organic matter is largely AOM, again consistent with relatively deepwater conditions. The comparatively organic nature of the sediments, indicated by their colour and a sudden increase in total organic carbon, is consistent with Rhodes et al. (1996)'s interpretation that this unit was formed under anoxic bottom water conditions. Anoxic bottom water conditions can be produced either when lake water is very deep, which does not appear to be the case here, or when the water is saline. The relatively low carbonate content allows siliceous microfossils (diatoms and chrysophyte cysts) to occur for the first time. The diatom assemblage is characterised by a mixture of mesosaline (*Nitzschia* spp., *Fragilaria fasciculata*) and hypersaline (*Amphora holsatica*, *A. coffaeformis*, *Mastogloia pumila*, *M. aquilegia*) species. This assemblage, which appears to indicate saline and thus very shallow conditions, supports the idea that anoxic conditions were created by salinity (rather than extreme water depth). On the other hand, the diatom assemblage is difficult to reconcile with the lithological evidence of even moderately deep water conditions. We assume that it reflects the initially highly saline waters produced by inputs of halite into the lake through salt-crust dissolution as the lake level rose.

The overlying unit (2.05-1.75m) is a grey, carbonate-rich, non-laminated lacustrine clay. The halite content is low and as a result of this, bottom water conditions are oxic and the organic content becomes reduced. The increase in carbonate content results in the disappearance of diatoms. There is no indication that the changes in the appearance of the sediments are related to changes in water depth. A sample from the base of this unit (2.035-2.05m) has been radiocarbon dated to 1905±50 yr B.P., while a sample from the base of the overlying unit (1.713-1.728m) has been radiocarbon dated to 1140±50 yr B.P. On the basis of these dates, this unit was deposited between 1910-1210 yr B.P.

The overlying unit (1.75-1.67m) is dark-coloured, sapropelic mud, similar to the unit between 2.40-2.05m. The change in sedimentation coincides with a peak in halite and is assumed to reflect a return to anoxic conditions consequent on the increased salinity of the water caused by a sudden input of halite into the system. The diatom assemblage is fully consistent with hypersaline conditions. There is no obvious reason why there should be a sudden flush of halite into the system. There is no indication that the changes in the nature of the sediment are related to changes in water depth.

The uppermost sediments (1.67-0m) are evaporite deposits and show a progressive change from mirabilite-rich deposits through gypsum-thenardite to pure halite upcore. The presence of this sequence of evaporites indicates progressive concentration of the lake waters as the lake dried out. A radiocarbon sample from within the unit (1.52-1.54m) has been dated to 330±80 yr B.P. Interpolation between this date and the date on an underlying unit suggests that desiccation of the lake started ca 925 yr B.P. Evaporite crusts normally form rather quickly, and this estimate of the age of the base of the evaporites therefore seems rather old.

In the status coding, very low (1) is indicated by evaporite deposits; low (2) by units representing concentrations of fluvial material deposited within the lake; moderately low (3) by playa lake deposits; intermediate (4) by lacustrine, non-laminated, silty clays and clays; high (5) by poorly-laminated lacustrine clays; and very high (6) by laminated lacustrine clays.



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Radiocarbon dates

H-693	37,800±1500	4.684m, carbonate, LM III Core
H-644	32,100±750	4.664-4.684m, carbonate, LM II Core, AMS date
H-689	10,120±100	3.979-3.999m, carbonate, LM II Core, AMS date
H-566	10,030±560	4.005-4.015m, organics, LM II Core, not used for chronology (reversal), AMS date
H-605	7210±100	3.596-3.616m, carbonate, LM II Core, AMS date
	5310± 95	Carbonate, from exposed lacustrine sequence 20 km southwest of modern salt
H-690	4500±80	3.244-3.264m, carbonate, LM II Core, AMS date
H-604	3440±120	2.82-2.84m, carbonate, LM II Core, AMS date
H-601	1905±50	2.035-2.05m, organics, LM II Core, AMS date
H-602	1140±50	1.713-1.728m, organics, LM II Core, AMS date
H-603	330±80	1.52-1.54m, organics, LM I Core

Coding

pre 32,000yr B.P.	intermediate (4)
32,000-11,600 yr B.P.	moderately low (3); NOTE dating extremely uncertain
11,600-10,350 yr B.P.	intermediate (4)
10,350-10,100 yr B.P.	low (2)
10,120-9160 yr B.P.	intermediate (4)
9160-7930 yr B.P.	very high (6)
7930-5630 yr B.P.	high (5)
5630-4250 yr B.P.	low (2)
4250-3460 yr B.P.	very high (6)
3460-3380 yr B.P.	low (2)
3380-3260 yr B.P.	very high (5)
3260-2780 yr B.P.	intermediate (4)
2780-2580 yr B.P.	low (2)
2580-350 yr B.P.	intermediate (4)
925-0 yr B.P.	very low (1)

Preliminary coding: 11-1-1999
Second coding: 25-04-2000
Final coding: 27-07-2000
Coded by: GY and SPH

3.29. Wulukekule Lake, Xinjiang Autonomous Region

Wulukekule Lake (35.67°N, 81.62°E, 4687m above sea level) is a closed-basin lake in the Ashikule Basin, in the middle part of the Kunlun Mountains, a depopulated zone on the northern Tibetan Plateau (Li, 1992). The Ashikule Basin originated through faulting. The Ashikule Basin has an area of 740km^2 (Li, 1992). Volcanic activity during the Quaternary has resulted in the creation of a number of sub-basins which are occupies today by hydrologically-independent lakes. The Wulukekule Lake occupied one of these sub-basins. The bedrock in the lake basin is igneous. A few seasonal streams enter the lake from the south. The lake water is supplied mainly by runoff and snow meltwater from the southern slopes of the Ashikule Basin (Li and Zhang, 1991). The lake area was 15.5km^2 in 1970, but decreased to 15km^2 by 1986 due to a long-term trend towards increasing aridity (Li, 1992). The water is saline and the pH value is 9.22 (Li, 1992). The climate in the catchment area is cold (-5 to -6°C mean annual temperature) and dry (100 to 300mm total annual precipitation).

A lacustrine terrace to the west of the modern lake provides evidence of a former high stand during the early- and mid-Holocene. The elevation of the top of the terrace is 4691.4 m a.s.l. A 3m-deep sediment profile (Profile C in Li, 1992; Profile D in Li and Zhang, 1991) from this terrace has been studied. Changes in relative water depth are reconstructed on the basis of changes in lithology and aquatic plant macrofossils from this profile (Li and Zhang, 1991; Li, 1992).

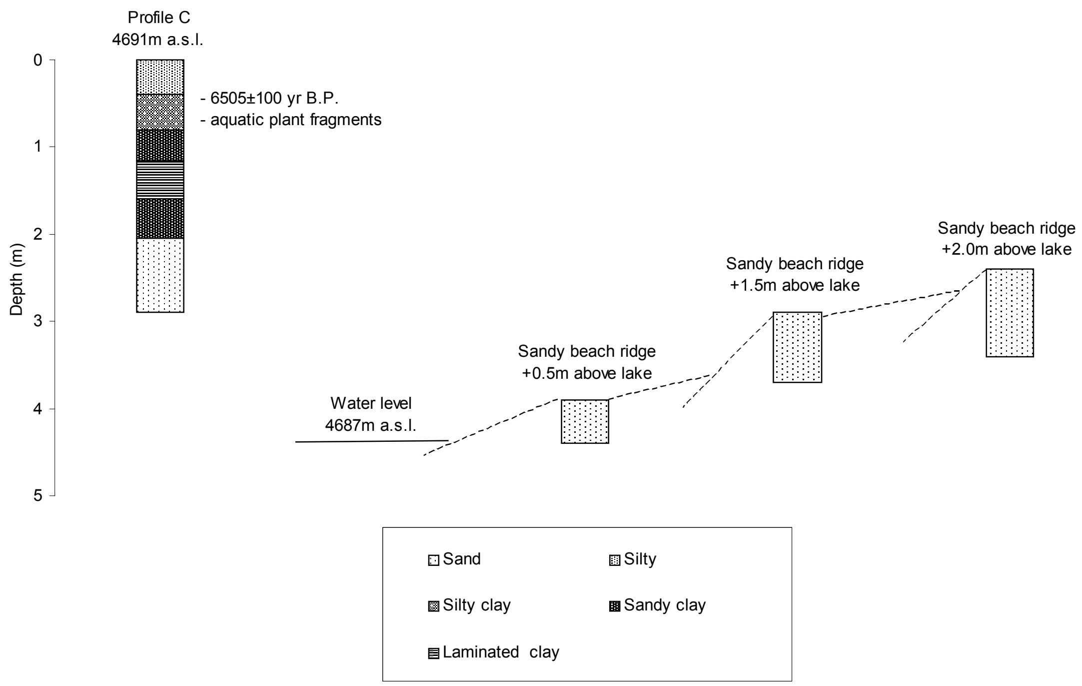
There is only one radiocarbon date from Profile C. A sample of aquatic material from 0.4m was radiocarbon-dated to 6506±100 yr B.P. This date indicates that the terrace was formed during a high stand prior to ca 6500 yr B.P. At this stage, the lake would have had a minimum level of ca 4691 m a.s.l. and a minimum area of 18km^2, ca 3km^2 larger than today.

The basal unit in Profile C (below 2.05m) is sand with crossbeds and diabeds (inclined < 10° bedding of strata). This unit is interpreted as a fluvial deposit (Li, 1992), and indicates that the lake level was below 4689m at this time. The overlying sediment (2.05-1.60m) is lacustrine sandy clay, indicating nearshore deposition (Li, 1992). The lake level was probably at ca 4690 m a.s.l. at this time. The overlying sediment (1.60-1.15m) is lacustrine clay with thin beds, characteristic of a deep-water deposit (Li, 1992) and thus indicating increased water depth. A return to non-bedded lacustrine sandy clay between 1.15-0.8m indicates decreased water depth. The overlying sediment (0.8-0.4m) is lacustrine silty clay, suggesting a slight increase in water depth. The unit contains aquatic plants and plant fragments, consistent with nearshore deposition. The uppermost sediment (0-0.4m) is silt, and is interpreted as a lake-beach deposit (Li, 1992).

Li (1992) has attempted to date the various lake-level fluctuations indicated by the changing lithology of the units within the profile by using the sedimentation rate calculated between two radiocarbon dates on a sedimentary sequence covering the deglaciation (ca 16,000 to 11,000 yr B.P.) from the nearby Ashikule Lake basin, and applying this sedimentation rate to estimate back from the radiocarbon date on Profile C. The Ashikule Lake is located within the same region, and only 5km from Wulukekule Lake, and Li (1992) presumably assumes that the transference of sedimentation rates from one basin to the other is justifiable since the two lake systems should show a similar sedimentary history in response to regional climate changes. However, there is no record that the Ashikule Lake underwent the significant changes during the Holocene that are recorded by the terrace in the Wulukekule Lake basin. Similarly, there is no evidence for a high lake stand during the deglaciation in the Wulukekule Lake basin similar to that observed in the Ashikule Lake basin. Furthermore, the Ashikule Lake derives most of its water from the northern slopes of the Ashikule structural basin, while Wulukekule derives its water from the southern slopes of this structural basin. It is therefore possible that the two lakes may have dissimilar histories since they have different hydrological source areas. We have therefore not adopted the chronology of lake-level changes given by Li (1992).

There are 3-5 sandy beach-ridges to the northeast of the lake. The highest ridge is at 1.8-2.0m and the lowest at 0.5m above modern lake level. Unfortunately these features are undated. They are thought to indicate lake-level lowering of 2m during the last few thousand years (Li, 1992).

In the status coding, low (1) is indicated the modern lake level at 4687 m a.s.l. and the undated sandy beach ridges between +0.5 and +2m above modern lake level; high (2) by the lacustrine terrace at +4.4m above modern lake level. A more detailed status coding, drawing on the lithological changes shown in Profile C, is possible but seems unwarranted given the lack of dating control.



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Radiocarbon date

6505±77	0.4m, aquatic plant, Profile C

(The sample was dated in ^{14}C Lab, Geography Institute of Chinese Academy of Science)

Coding

7000-6500 yr B.P.	high (2)
0 yr B.P.	low (1)

Preliminary coding: 4-12-1998
Second coding: 13-12-1998
Final coding: 21-12-1998
Coded by GY and SPH

3.30. Xiaoshazi Lake, Xinjiang Autonomous Region

Xiaoshazi Lake (36.97°N, 90.73° E, 4106m above sea level) is a fresh-water lake in the Kumuku Basin, in the middle part of the Kunlun Mountains. The basin lies in a depopulated region on the northern Tibetan Plateau (Li, 1992). The Kumuku Basin is a large structural basin, formed by faulting. Long-term erosion within the Kumuku Basin has resulted in the creation of several inset sub-basins, including the Xiaoshazi and Beilikekule basins. There is no information about the area of the Xiaoshazi Lake catchment basin, but the Kumuku Basin area is 45,000km^2 (Huang et al., 1996). A stream enters the Xiaoshazi Lake from the southwest, and an outlet stream goes northwestwards to Dashazi Lake, which is closed. The area of Lake Xiaoshazi was 33km^2 in 1978 but was reduced to 25km^2 by 1986 due to the long-term trend towards more arid climate (Li, 1992). The lake water is supplied mainly by runoff and snow meltwater from the basin (Li and Zhang, 1991). The water salinity is 0.323 g/L and the pH value is 8.1 (Li, 1992). The Kumuku Basin is characterised by alpine desert vegetation, with Chenopodiaceae and dominated by *Ceratoides* (Huang et al., 1996). The climate in the basin is cold (-5 to -6°C mean annual temperature) and dry (100 to 300mm total annual precipitation) (Huang et al., 1996).

Two lacustrine terraces, one 20m and the other 8m above modern lake level, provide evidence of former high stands. A sediment profile was taken from the 20m high-terrace, ca 2km east to the lake margin (36.79°N, 90.93°E; called Profile A in Li and Zhang, 1991; and Profile G in Li, 1992). A second profile was taken from the 8m low-terrace, ca 0.5km away from the lake margin. These profiles provide sedimentary records back to ca 11,000 yr B.P. (Li and Zhang, 1991; Li, 1992). The preserved shorelines provide a basis to estimate the palaeolake areas at specific times (Li, 1992). Changes in relative water depth are reconstructed on the basis of changes in lithology, geochemistry, diatom assemblages and aquatic plants within the profiles (Li and Zhang, 1991; Li, 1992). The chronology is based on two radiocarbon dates (both older than 8000 yr B.P.) from Profile A (Li and Zhang, 1991).

The basal sediments in Profile A (2.4-1.6m) are laminated lacustrine clay, indicating deep water at the site. The unit has pH values of 8.4-8.1 and organic contents of 1.4-1.7%. Li and Zhang (1991) interpreted these values as typical for lakes at very high altitudes and under cold conditions. The diatom assemblage is dominated by planktonic *Cyclotella* spp. with epiphytic or littoral forms tolerant of fresh to brackish water including *Cocconeis placentula*, *Nitzschia denticula*, and *Amphora mexicana*, consistent with deep water. The unit was formed between ca 13,000-10,730 yr B.P. by extrapolation of the sedimentation rate (0.0257 yr/cm) between the two radiocarbon dates from 1.5m and 0.9m.

A thin layer of aquatic plants occurs at ca 1.4-1.6m, indicating shallower conditions. A sample of these aquatic plants from 1.5m was radiocarbon dated 10,693±238 yr B.P., suggesting this shallow water phase occurred between ca 10,655-10,730 yr B.P. using the sedimentation rate (0.0257 cm/yr) between the radiocarbon dates from the profile.

The overlying sediment (1.4-1.0m) is laminated lacustrine clay, indicating a return to deep water at the site. The unit is dated to ca 10,655-8395 yr B.P. The diatom assemblage is dominated by planktonic *Cyclotella* spp., consistent with deep water.

The overlying sediment (1.0-0.8m) is a thin layer of black aquatic plants, indicating a return to shallower conditions. A sample of aquatics at 0.9m was radiocarbon-dated to 8356±172 yr B.P., suggesting the shallow water phase occurred between 8395-8315 yr B.P.

The overlying sediment (0.8-0.4m) is laminated lacustrine clay, indicating a return to deep water at the site. The diatom assemblage is dominated by planktonic *Cyclotella* spp., consistent with deep water. The unit is dated to ca 8315-6410 yr B.P., by extrapolation of the sedimentation rate between the two radiocarbon dates from the section.

Huang et al. (1996) correlated the pollen assemblage from 0.8-0.4m which is characterised by high percentages of *Artemisia* and low percentages of Chenopodiaceae, with a similar assemblage at the nearby site of Aqigekule. On the basis of two radiocarbon dates on lacustrine material at Aqigekule, Huang et al. (1996) suggest the change to laminated clay deposition occurred between ca 6000-7000 yr B.P. However, the chronology at Aqigekule is not well-constrained since the two dates come from different lacustrine units separated by a period of aeolian deposition. Thus, there does not seem to be a good justification for changing the chronology based on the actual dates on the Xiaoshazi profile.

The uppermost sediment (0-0.4m) in Xiaoshazi profile is laminated lacustrine sandy clay. The presence of sand suggests decreased water depth. The unit has pH values of 8.7-8.9 and an organic content <1.17%. The diatom assemblage is dominated by *Anomeoeneis sphaerophora* and *Navicula oblonga* (benthic/epiphytic forms). The decreased abundance of *Cyclotella* spp. and the increased abundance of benthic/epiphytic diatoms is consistent with decreased water depth. By extrapolation of the sedimentation rate on the underlying sediment, this unit was formed between ca 6410-4850 yr B.P. Li (1992) argued that this lacustrine unit, the top of which has an elevation of ca 4126 m a.s.l., represents a high lake stand, ca 20m above the modern lake level. He also estimated the lake area was ca 35km^2 on the basis of intermittent shorelines along ca 4125m contour around the basin. The occurrence of a laminated unit at the top of terrace profile implies that the lake level was significantly higher than at 4126m. Thus Li's estimate is a minimum estimate of the lake depth and area.

The 8m terrace surface at 4114 m a.s.l. is well preserved and 700-1000m wide, and marks a lake stand 8m higher than today. The profile from the 8m-terrace consists of 3m of lacustrine clay. Unfortunately, these sediments are undated, so there is no way of estimating when this second high stand occurred. Modern lake level is the lowest in the lake history.

In the status coding, low (1) is indicated by modern lake level at 4106m; intermediate (2) by aquatic plant layers between lacustrine sediments from the +20m terrace; moderately high (3) by lacustrine laminated sandy clay with a mixed planktonic/benthic/epiphytic diatom assemblage; and high (4) by lacustrine laminated clay with a diatom assemblage dominated by planktonics.

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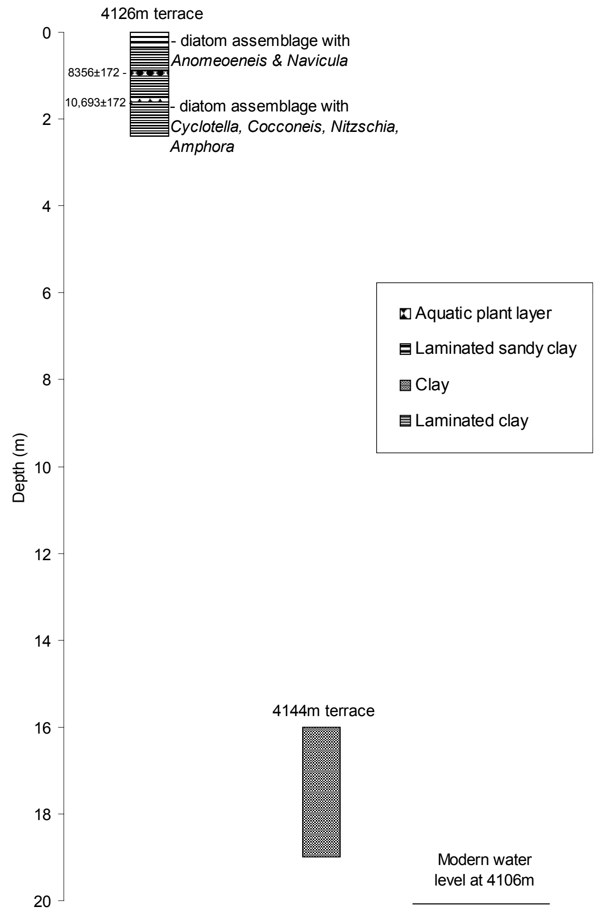
Li SK (1992) Fluctuation of closed lake-level and its climatic significance on the middle Kunlun Mountains. Journal of Lake Sciences 4(1): 19-30 (in Chinese)



Radiocarbon dates

10,693±238	1.5m, aquatic plants, Profile A
8356±172	0.9m, aquatic plants, Profile A

(The samples were dated in ^{14}C Lab, Geography Institute of Chinese Academy of Science)

Coding

ca 13,000-10,730 yr B.P.	high (4)
ca 10,730-10,655 yr B.P.	intermediate (2)
ca 10,655-8395 yr B.P.	high (4)
ca 8395-8315 yr B.P.	intermediate (2)
ca 8315-6410 yr B.P.	high (4)
ca 6410-4850 yr B.P.	moderately high (3)
0 yr B.P.	low (1)

Preliminary coding: 30-11-1998
Second coding: 18-12-1998
Third coding: 31-12-1998
Final coding: 27-3-1999
Coded by GY and SPH

3.31. Akesaiqin Lake, Xizang (Tibet) Autonomous Region

Akesaiqin Lake (in standard Chinese phonetics; Aksayqin also given in Li SJ et al., 1991 and Aksaichin in Times World Atlas, 1967) (35.20°N, 79.83° E, 4840m above sea level) is a large salt lake in the West Kunlun Mountains, on the western Tibetan Plateau. The lake lies in a large structural basin, with an area of ca 18,800 km^2 and a threshold elevation of 4935 m a.s.l. (Wang et al., 1990). This structural basin contains a number of inset lacustrine basins, including those of the modern lakes Guozacuo, Akesaiqin, Tianshuihai, North Tianshuihai and Kushiuhai. The inset basins originated through faulting. Although tectonism is still active in this region, the effects of uplift on the lake system is thought to be small compared to the effects of climate change (Wang et al., 1990).

During the early Quaternary, the Akesaiqin basin was part of a much larger palaeolake, which unified the modern lakes of Akesaiqin, Tianshuihai, North Tianshuihai and Kushiuhai (Wang et al., 1990; Li BY, 1991). The extent of the mega-palaeolake is defined by intermittent shorelines between 4880-4890 m a.s.l. and is estimated to have been ca 2650 km^2 (Wang et al., 1990) and the catchment area was equivalent to the whole of the structural basin (i.e. ca 18,800 km^2). However, Guozacuo lake does not appear to have been connected with the mega-palaeolake at any time. On the basis of a single radiocarbon date, with an age >45,000 yr B.P., Wang et al. (1990) suggest that the mega-palaeolake existed before 45,000 yr B.P. The mega-palaeolake was fresh and had an overflow to the Karakala River.

As climate became more arid after ca 45,000 yr B.P., lake level fell and the individual basins became distinct. Akesaiqin Lake was originally connected to Tianshuihai Lake via a river channel, but by the early Holocene the level of Akesaiqin Lake had dropped below 4845 m a.s.l. and this connection was broken (Wang et al., 1990). Tianshuihai (4830 m a.s.l.) and North Tianshuihai (4800 m a.s.l.) lakes are still part of the same system, since they both have an intermittent overflow to Kushuihai Lake (4754 m a.s.l.). Akesaiqin Lake (4840 m a.s.l.) lies 65km southeast of North Tianshuihai Lake, and ca 40km east-southeast of Tianshuihai. There is no surface or subsurface connection between Akesaiqin Lake and the other lakes in the structural basin. Its history over the last 45,000 yr B.P. can therefore be considered separately from that of the other lakes in the basin.

When Sven Hedin (1922) investigated the Akesaiqin Lake basin in the early part of this century, he called it the Asian White Desert because the lake was nearly dry. Today, Akesaiqin Lake has an area of 160km^2 (estimated as 158km^2 and 192km^2 in Wang HD et al., 1987 and Wang FB et al., 1990 respectively; the area of 160km^2 is from the latest publication by Li SJ et al., 1991) and a maximum depth of 12.6m. The lake water is supplied by runoff and snow/glacier meltwater from high mountains in the catchment. A stream from the south slope of the Kunlun Mountains enters the west side of the lake. The water salinity is 54.26 g/L (Li SJ et al., 1991). The sediments deposited in the lake bottom today are brown-yellow sandy clay (Wang et al., 1990).The bedrock in the lake basin is Jurassic-Cretaceous conglomerate and sandstone (Wang et al., 1990). The climate in the lowland part of the basin is extremely cold and dry, with -5°C mean annual temperature and 20-40mm total precipitation (Wang et al., 1990) but 2500mm annual evaporation (Li et al., 1994). Higher precipitation levels (ca 400-500mm) are recorded in the Kunlun Mountains, which form the northern margin of the structural basin (Li BY et al., 1991).

There are fragments of lacustrine terraces at elevations of 4845m (+5m above modern lake level) and 4850 m a.s.l. (+10m above modern lake level), intermittently located to the west, north and east of the lake. However, investigations of sediment profiles indicate that fragments from approximately the same elevation were not necessarily formed at the same time. This is consistent with the fact that the terraces appear to be largely erosional remnants of lake-bottom sediments. Three sediment profiles (NW and NE both at 4850m, and Profile N at 4845m) provide sediment records back to ca 35,000 yr B.P. (Wang et al., 1990; Li SJ et al., 1991). Changes in lake status are reconstructed on the basis of changes in geomorphology, lithology and aquatic plants, and follow the interpretations of Wang et al. (1990) and Li SJ et al (1991). The chronology is based on 8 radiocarbon-dates from the profiles.

Profile NW, the top of which is at an elevation of 4850m, is 10-m deep. The basal sediment (9-10m) is grey laminated lacustrine clay, suggesting a deepwater lake. Two samples from ca 9m and ca 10m of the laminated clay unit were radiocarbon-dated to 33,065±585 and 34,735±820 yr B.P., suggesting the deep-water lake was formed between 33,000-35,000 yr B.P. The lake level must have been > 4842m and given the laminated nature of the sediments was likely to have been considerably higher. The lake area was larger than today (Wang et al., 1990; Li SJ et al., 1991). The overlying unit is lacustrine sand (8-9m), characteristic of nearshore deposition and indicates that lake level dropped to ca 4842m after ca 33,000 yr B.P. The uppermost 8m (0-8m) is gravel and sand which originated as lake beach deposits. The deposition of 8m of beach deposits probably indicates that the lake level was gradually rising during this interval.

The 3-m deep Profile NE, the top of which is at an elevation of 4850m, consists of lacustrine sediments. The basal sediments (2.5-3.0m) are grey clay, suggesting a moderately deep lake. This unit can be dated to between 24,210-22,520 yr B.P., by extrapolation of the sedimentation rate (0.03125 cm/yr) between two samples from 2.50m and 1.25m which were radiocarbon-dated to 22,520±690 and 18,520±305 yr B.P. respectively. The lake level must have been > 4848m during this moderately deepwater phase, and may have been somewhat higher. However, the sediments are not laminated and so the lake was not as deep as during the interval between 35,000 and 33,000 yr B.P. The overlying sediment in Profile NE (2.5-2.4) is a layer of aquatic plant remains, suggesting decreased water depth after 22,520 yr B.P. The presence of aquatic plant remains suggests nearshore deposition and indicates a lake level of ca 4848 m a.s.l. The overlying sediment (2.4-2.1m) is grey clay, suggesting a return to deep water conditions between 22,200-21,570 yr B.P. The overlying sediment (2.1-1.8m) is a layer of aquatic plant remains, suggesting decreased water depth after 21,570 yr B.P. The overlying sediment (1.8-1.5m) is laminated black clay, suggesting increased water depth between 20,300-19,340 yr B.P. The disappearance of aquatic plant remains is consistent with deep water. The presence of laminations indicates that the water level was considerably higher than it was between e.g. 24,210-22,520 yr B.P. and may have been as high as during the interval between 35,000 and 33,000 yr B.P. The overlying sediment (1.5-0.2m) is non-laminated clay, suggesting decreased water depth after 19,345 yr B.P. The upper part (1.0-0.2m) of this unit contains abundant aquatic plants, suggesting further shallowing between 18,025-16,450 yr B.P. The uppermost sediment in Profile NE (above 0.2m) is silt, suggesting a further decrease in water depth after 16,450 yr B.P. A sample at 0.1m was radiocarbon-dated to 16,235±120 yr B.P.

Profile N, the top of which has an elevation of 4845m, consists of 5m of lacustrine grey silty clay with discontinuous thin beds. The lithology indicates moderately deep water conditions. The unit contains only a few aquatic plant remains, though including *Potamogeton* spp., consistent with moderately deep water. Three samples at 0.1m, 2.6m and 2.8m were radiocarbon-dated to 13,920±400, 15,720±200 and 15,960±240 yr B.P. respectively, suggesting this lacustrine phase occurred between 13,900-18,600 yr B.P. The lake level must have been > 4845 m a.s.l. and may have been somewhat higher. It is plausible that this unit represents the distal facies of the non-laminated clay and silt found at the top of Profile NE.

There are four gravel-sand ridges in the northern lowlands of the lake (Wang et al., 1990), indicating a series of former high lake stands. Unfortunately none of them are dated. The highest ridge is +25m above the modern lake level (Wang et al., 1990) and the other ridges are at elevations a few meters lower. Wang et al. (1990) estimated they were formed between ca 16,000-13,000 yr B.P. by correlation with radiocarbon-dated shorelines from nearby basins in the region. However, it seems more likely that these ridges correspond to shoreline positions during the glacial maximum. Although there is a record of lacustrine deposition between 16,000 and 13,000 yr B.P. in the N Profile, the sediments are non-laminated and the lake level was unlikely to have been significantly higher than say 4850 m a.s.l. The laminated sediments in Profile NE which occur at an elevation of 4848 and 4848.5 m a.s.l., must have been deposited in a water depth of at least ca 10m and thus require that the lake level was > 4858m (i.e. > +18m above modern lake level and possibly therefore at an elevation close to the +25m indicated by the gravel-sand ridges) between ca 20,300-19,340 yr B.P.

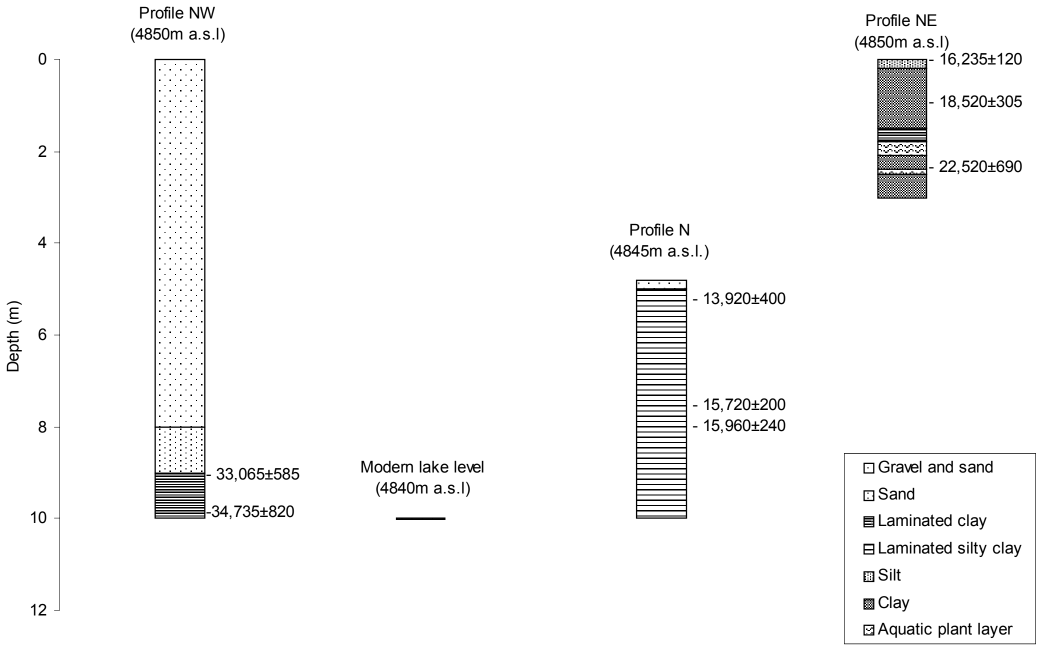
Geomorphological analysis suggests that the modern lake is smaller than at any time during its history (Wang et al., 1990).

In the status coding, very low (1) is indicated by modern lake level; low (2) by nearshore sand deposition or beach deposition in Profile NW; moderately low (3) by aquatic plant layers, indicating lake level around ca 4848 m a.s.l. in Profile NE; intermediate (4) by silt deposition in Profile NE and non-laminated clay deposition in Profile N; moderately high (5) by non-laminated clays with abundant aquatics in Profile NE; high (6) by non-laminated clay deposition in Profile NE; very high (7) by laminated clay deposition in Profiles NE and NW.

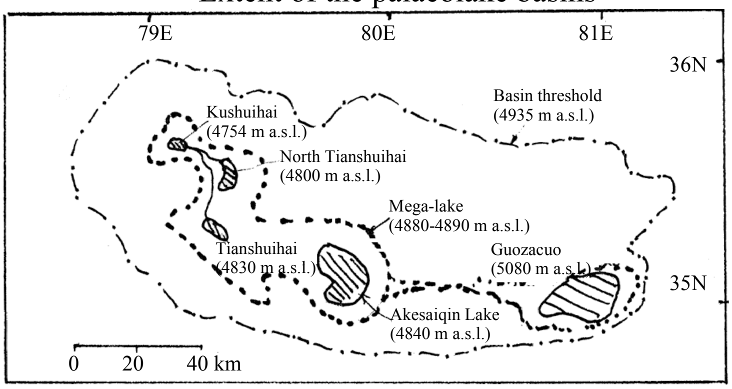
Radiocarbon dates

34,735±820	10m, clay, Profile N *
33,065±585	9m, clay, Profile N
22,520±690	2.5m, aquatic remains, Profile NE
18,520±305	1.25m, clay, Profile NE *
16,235±120	0.1m, silt, Profile NE *
15,960±240	2.8m, silty clay, Profile N
15,720±200	2.6m, silty clay, Profile N
13,920±400	0.1m, silty clay, Profile N

The 3 samples marked with * were dated in ^{14}C Lab, Department of Geography, Lanzhou University; the rest in ^{14}C Lab, Geo and Ocean Science, Nanjing University



Extent of the palaeolake basins



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Coding

35,000-33,000 yr B.P.	very high (7)
33,000-? yr B.P.	low (2)
? -24,210 yr B.P.	not coded (no record)
24,210-22,520 yr B.P.	high (6)
22,520-22,200 yr B.P.	moderately low (3)
22,200-21,570 yr B.P.	high (6)
21,570-20,300 yr B.P.	moderately low (3)
20,300-19,340 yr B.P.	very high (7)
19,340-18,025 yr B.P.	high (6)
18,025-16,450 yr B.P.	moderately high (5)
16,450-13,900 yr B.P.	intermediate (4)
13,900-100 yr B.P.	not coded (no record)
0 yr B.P.	very low (1)

Preliminary coding: 25-11-1998
Second coding: 20-02-1999
Final coding: 27-03-1999
Coded by GY and SPH

3.32. Bangge Lake, Xizang (Tibet) Autonomous Region

Bangge Lake (given as Bangkog in Zheng et al., 1989; 31.75°N, 89.57°E, 4520m above sea level) is a salt lake in the interior part of the Tibetan Plateau. Bangge Lake consists of three sub-basins (called Lakes I, II and III) joined by narrow channels (Zheng et al., 1989). Lake I is the upstream sub-basin and has a water-surface elevation of 4525 m a.s.l. There is no surface input into Lake I. Lake I discharges into Lake II (surface water elevation 4522 m a.s.l.) which in turn discharges into Lake III (surface water elevation 4520 m a.s.l.). A small stream discharges into Lake III from the southeast. This is the only surface input to the lakes, which are otherwise fed by direct precipitation and springs. The lakes have water depths between 0.3m (Lake I) and 1m (Lake III). Lake I and II are seasonally dry, however Lake III is a permanent water body. At their maximum (summer) extent, Lakes I, II and III have areas of 5.4, 50 and 80km^2 respectively. The lake water is saline. The salt content of the water in Lakes I, II and III respectively is 168.7, 235-258 and 189-403 g/L and the pH is 9-9.56, 9 and 9-10.2 (Zheng et al., 1989). The modern sediments consist of mirabilite-bearing carbonate clay in the centre of the lake and mirabilite-bearing carbonate sand with gravels in the nearshore zone (Zheng et al., 1989). The three saline lakes are inset within a playa. The existence of this playa indicates that Bangge Lake was larger in the past. The Bangge Lake basin is surrounded by high mountains (maximum elevation of 6440 m a.s.l.). The Bangge Basin originated through faulting in the Mesozoic. The underlying bedrock is Cretaceous conglomerate, sandstone and argillite. The climate in the catchment is cold (annual mean temperature of 1.5°C) and arid (total annual precipitation of 308.3mm, 2238.6mm total annual evaporation) (Zheng et al., 1989).

Three major surfaces (called terraces by Zheng et al., 1989) have been recognised within the basin. The uppermost of these (T3) occurs as a sloping surface between +100m and +150m above modern lake level and was created by erosion (i.e. is probably a planation surface). The second terrace (T2) is probably a constructional feature, which occurs as a sloping surface between +10 and +80m above modern lake level and at its upper end abuts onto T3. The lowermost terrace (T1) extends from the modern lake beach to an elevation of ca +10m above modern lake level and abuts onto T2. T1 is also thought to be a constructional feature. Both of the lower terraces are covered by sequences of sub-parallel beach ridges. There are at least 29 such beach ridges on the T2 surface and 21 beach ridges on the T1 surface.

The structure of the lower, constructional terraces has been investigated by means of a number of drill cores and geological sections (Zheng et al., 1989). The geomorphic position of four of the profiles described by Zheng et al. (1989) cannot be located. However, there is a single 3.57m-long profile from Terrace II (Profile II-4), taken from the northeast of Lake I. The top of the profile has an elevation of +55m above modern lake level in Lake I (4580 m a.s.l.). The profile penetrates lacustrine deposits but is undated. The deposits are assumed to be Middle Pleistocene in age. There are three cores/sections from Terrace 1 (Core B1, Core CK2 and the Bangkog Lake II terrace profile).

Core B1 was taken from the T1 terrace to the west of Lake III. The elevation of the top of the core is not given, but we estimate that it is ca 3-5m above modern lake level (4525 m a.s.l.). Core B1 is 959m long. The deposits below 805m are Cretaceous in age. The remainder of the core provides a record of quasi-continuous lacustrine deposition in the Bangge basin from the Miocene to the Late Quaternary (Zheng et al., 1989). The uppermost sediments (lacustrine beach sands and gravels), although undated, have been assigned to the last glaciation. There are no Holocene sediments in this core. The presence of lacustrine beach deposits at elevations above the modern lake level attest to higher-than-present lake levels during the glacial.

A 11.8m-long core (Core CK2) taken from Terrace T1 at the margin of Lake II provides a sedimentary record back to ca 20,000 yr B.P. (Zheng et al., 1989). The elevation of the top of this core is ca 4525 m a.s.l. Changes in water depth and salinity since ca 20,000 yr B.P. can be reconstructed from changes in the lithology and sediment geochemistry of this core. Geochemical analyses show changes in the relative abundance of magnesite ($MgCO_3$) and calcite ($CaCO_3$). Magnesite is generally formed under high alkalinity and high pH values (9-12) while calcite is formed at lower alkalinity and pH values (Zheng et al., 1989). We therefore interpret high relative levels of magnesite as indicating shallower conditions and high relative calcite as indicating deeper conditions. The chronology of changes in water depth is based on 4 radiocarbon dates from Core CK2.

The basal sediment in Core CK2 (11.8-11.6m) is pale-yellow, lacustrine carbonate clay. The unit contains < 5% $MgCO_3$ and ca 6% $CaCO_3$. The fine-grained nature of these sediments and the relative abundance of calcite indicate that the lake was fresh and moderately deep. The overlying unit (11.6-11.48m) is pale-yellow, clayey fine sand and gravel. The change in lithology indicates shallower conditions than formerly. There is no apparent change in the geochemistry (< 5% $MgCO_3$ and 6% $CaCO_3$). The overlying unit (11.48-11.27m) is pale-yellow, lacustrine carbonate clay, suggesting a return to deep water. The values of $MgCO_3$ remain low (< 5%) but $CaCO_3$ increases to 20%. This change in the geochemistry indicates a considerably freshening of the water and is consistent with increased water depth. By extrapolation of the sedimentation rate (0.0264 cm/yr) between radiocarbon dates in overlying units, this unit was formed sometime before ca 20,100 yr B.P. These relatively deep-water deposits can be correlated with the glacial-age lacustrine beach sands and gravels found at the top of Core B1. This suggests that the lake level was ca 4525 m a.s.l. and that the maximum depth of the lake at the site of Core CK2 was ca 11-12m.

The overlying unit (11.27-7.9m) in Core CK2 is black, mirabilite- or thenardite-bearing, carbonate muddy clay. The presence of mirabilite indicates saline water conditions. The organic content of the mud, indicated by its colour, is consistent with moderately shallow water. The unit can be subdivided on the basis of changes in geochemistry. Between 11.27-10.7m, the unit contains ca 5% $MgCO_3$ and ca 20% $CaCO_3$. Between 10.7-9.8m, the unit contains 10-40% $MgCO_3$ and ca 10% $CaCO_3$. Between 9.8-9.0m, the unit contains 40-60% $MgCO_3$ and ca 5% $CaCO_3$, while sediments between 9.0-8.61m contain ca 5% $MgCO_3$ and ca 10% $CaCO_3$. The uppermost sediments (8.61-7.9m) contain ca 10-40% $MgCO_3$ and ca 3-6% $CaCO_3$. This sequence shows a gradual increase in salinity up to 9.0m, a return to fresher conditions and then somewhat more saline conditions afterwards. A sample from 10.2m was radiocarbon-dated to 16,800±210 yr B.P. and a sample from 8.55m was radiocarbon-dated to 9795±115 yr B.P. Interpolation between these dates suggests that the first shallow but moderately fresh water stage occurred between 20,100-17,940 yr B.P. The second phase of shallow and increasingly saline water conditions occurred between 17,940-14,530 yr B.P. and the maximum shallowing occurred between 14,530-11,500 yr B.P. The return to shallow but moderately fresh water conditions is dated to between 11,500-

10,020 yr B.P. and the final phase of moderately saline conditions occurred between 10,020 and 9440 yr B.P.

The overlying unit (7.9-7.4m) is black, mirabilite-bearing marl. The change in lithology and the organic nature of the sediments is consistent with shallowing relative to the underlying unit. The increase in $MgCO_3$ (to 50%) and decrease in $CaCO_3$ (to < 3%) is consistent with this interpretation. The unit is dated to 9440-9200 yr B.P.

The overlying unit (7.4-6.6m) is black, mirabilite-bearing carbonate muddy clay. The change to muddy clay deposits indicates somewhat increased water depth. A decrease in $MgCO_3$ to 20% and an increase in $CaCO_3$ to > 6% is consistent with freshening. This unit is dated to between 9200-8800 yr B.P.

The overlying unit (6.6-6.0m) is black, mirabilite-bearing marl, suggesting decreased water depth between 8800-8500 yr B.P. High levels of $MgCO_3$ (to 60%) and decreased $CaCO_3$ (to < 3%) are consistent with increasingly saline conditions.

The overlying unit (6.0-5.4m) is black mirabilite-bearing carbonate muddy clay, suggesting increased water depth between 8500-8200 yr B.P. The change in the relative importance of $MgCO_3$ (to < 10%) and $CaCO_3$ (to > 6%) is consistent with this interpretation.

The overlying unit (5.4-5.2m) is black mirabilite-bearing marl, suggesting decreased water depth between 8200-8100 yr B.P. High levels of $MgCO_3$ (to 60%) and decreased $CaCO_3$ (to < 3%) are consistent with increasingly saline conditions.

The overlying unit (5.2-4.58m) is black mirabilite-bearing carbonate muddy clay, suggesting increased water depth between 8100-7830 yr B.P. The relative abundance of $MgCO_3$ (ca 20%) and $CaCO_3$ (10%) is consistent with freshening.

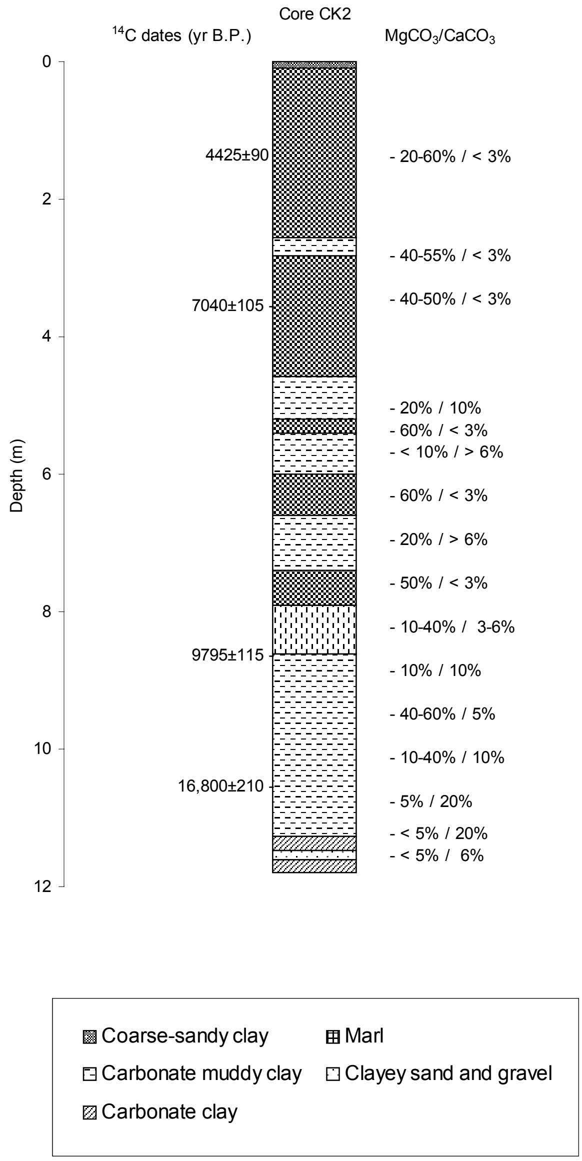
The overlying unit (4.58-2.82m) is black, mirabilite-bearing marl, suggesting decreased water depth. The increase in $MgCO_3$ (to ca 40-50%) and decrease in $CaCO_3$ (< 3%) is consistent with this interpretation. A sample from 3.0m was radiocarbon-dated to 7042±105 yr B.P. On the basis of interpolation between this date and over/underlying dates, this unit formed between 7830-6780 yr B.P.

The overlying unit (2.82-2.57m) is black, mirabilite-bearing carbonate muddy clay, suggesting increased water depth between 6780-6420 yr B.P. However, there is no significant change in the geochemistry.

The overlying unit (2.57-0.3m) is grey-white, mirabilite-bearing marl, suggesting decreased water depth after ca 6420 yr B.P. $MgCO_3$ is always abundant (>20%, and reaching maximum levels of 60%) and the $CaCO_3$ content is always low (<3%) within this unit, consistent with shallow conditions. Variations in the amount of $MgCO_3$ up-unit appear to reflect the presence/absence of other evaporites (including borate and sodium salts) rather than significant changes in water depth. A sample from 1.22m within this unit was radiocarbon-dated to 4425±30 yr B.P.

The uppermost sediment (0.3-0.1m) is pale-yellow, carbonate coarse-sandy clay. Zheng et al. (1989) interpret this unit as a nearshore deposition. This suggests very shallow water after 370 yr B.P.

In the status coding, very low (1) is indicated by coarse sand clay in Core CK2; low (2) by marl (with $MgCO_3$ values >20%; moderately low (3) by carbonate muddy clay with > 40% $MgCO_3$; intermediate (4) by carbonate muddy clay with 10-40% $MgCO_3$; high (5) by carbonate muddy clay with <10% $MgCO_3$; very high (6) by lacustrine carbonate clay with < 5% $MgCO_3$ in Core CK2 and beach deposits at elevations of 4525m in Core B1.



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Radiocarbon dates

16,800±210	11.40m, carbonate, Core CK2
9795±115	8.55m, marl, Core CK2
7040±105	3.0m, clay, Core CK2
4425±90	1.22m, clay, Core CK2

(The samples were dated by the ^{14}C Lab of Seismology and Geology Institute, the State Bureau of Seismology, China).

Coding

Pre 20,100 yr B.P.	very high (6)
20,100-17,940 yr B.P.	high (5)
17,940-14,530 yr B.P.	intermediate (4)
14,530-11,500 yr B.P.	moderately low (3)
11,500-10,020 yr B.P.	high (5)
10,020-9440 yr B.P.	intermediate (4)
9440-9200 yr B.P.	low (2)
9200-8800 yr B.P.	intermediate (4)
8800-8500 yr B.P.	low (2)
8500-8200 yr B.P.	high (5)
8200-8100 yr B.P.	low (2)
8100-7830 yr B.P.	intermediate (4)
7830-6780 yr B.P.	low (2)
6780-6420 yr B.P.	moderately low (3)
6420-370 yr B.P.	low (2)
370-0 yr B.P.	very low (1)

Preliminary coding: 02-02-1999
Second coding: 27-07-2000
Final coding: 30-07-2000
Coded by GY and SPH

3.33. Bangongcuo, Xizang (Tibet) Autonomous Region

Bangongcuo (also called Bangong Co; 33.67-33.73°N, 79.0-79.83°E, 4241m above sea level) is a large closed lake in a fault basin between the Kunlun Mountains and the Kalakunlun Mountains, on the western Tibetan Plateau. The basin was created by faulting during the Mesozoic and appears to have been tectonically stable during the Late Quaternary (Li BY et al., 1991). The lake is long (159km from east to west) and narrow (2.6km mean width from north to south) with an area of 412km^2 and a mean water depth of 18m (Wang et al., 1987). The lake is fed by a few rivers which drain the high mountains to the east of the basin (Li YF et al., 1991, 1994). These mountains reach elevations of 6200-6800 m a.s.l. and there is permanent snowcover above ca 6000 m a.s.l. There are no significant water inputs to the western part of the lake. As a result, there is a significant gradient in salinity from the eastern part (0.72 g/L) to the western part (19.61 g/L) of the basin. The modern climate in the basin is cold (0°C mean annual temperature) and dry (60.4mm total precipitation and 2465.3mm annual evaporation) (Li YF et al., 1991). The Bangongcuo basin is characterised by alpine desert vegetation, dominated by Chenopodiaceae, Gramineae, Compositae and Cyperaceae (Huang et al., 1996).

Geomorphological investigations have identified a number of terraces, created by erosion of lacustrine bottom deposits as lake levels decreased, to the east of the modern lake (Li BY et al., 1991). The highest of these terraces, the top of which is more than 100m above the modern lake level, was found 100km to the east of the lake. The terrace consists with grey-white lacustrine clay deposits. Li BY et al. (1991) estimated that this lake had an area of ca 970km^2 (ca 1.5 times of modern lake) and would have united the basins of Changmucuo, Abucuo and Aiyongcuo, which are hydrologically independent lakes today. Unfortunately this lake terrace is not dated.

A second lake terrace was found at Tagutuqiong, 30km to the east of the modern lake (Li YF et al., 1991; Huang et al., 1989). The elevation at the foot of the terrace is ca 4300 m a.s.l. and at the top is ca 4319.5 m a.s.l. (i.e. +78.5m above the modern lake level). The terrace sediments are exposed along a natural cutting over ca 150m horizontal distance and this has allowed a 19.5m composite profile (Profile I) from the bottom to the top of the Tagutuqiong terrace to be reconstructed. This composite profile covers the interval between ca 40,000-24,000 yr B.P. (Li YF et al., 1991; Huang et al., 1989). A second profile (Profile II) through the Tagutuqiong terrace, ca 20km away from the modern lake, provides a sedimentary record covering the interval between 40,000-18,000 yr B.P. (Li BY et al., 1991). A sample from the uppermost terrace sediments in Profile II has been radiocarbon dated to 18,187±167 yr B.P., indicating that the lake regression that created the Tagutqiong terrace occurred after ca 18,000 yr B.P.

The basal deposits in Tagutuqiong Profile I are grey lacustrine silty clay (0-2m from the base of the profile). The lithology suggests moderately deep water conditions at the terrace site. The ostracode assemblage is dominated by *Leucocythere postilirata* and *L. subculpta*, two species now living in fresh to semi-fresh water of Bangongcuo and tolerant of a wide range of shallow and muddy bottom conditions. Thus, the ostracode assemblage is consistent with relatively deep water conditions. On the basis of extrapolation of the sedimentation rate (0.2434 cm/yr) between radiocarbon dates from a depth of 2.25m and 11.6m, the basal unit was formed before ca 39,600 yr B.P.

The overlying unit (2.0-2.25m) is an incompletely decomposed peat with abundant visible plant remains, mostly of aquatic plants. This so-called "grass peat" (Huang et al., 1989) deposit is characteristic of shallow water, marshy conditions. There is no aquatic pollen in the unit. The ostracode assemblage is characterised by *Limnocythere dubiosa*, a species characteristic of fresh to slightly saline water (< 34 g/L). This change in the ostracode assemblage is consistent with the inference of shallower conditions. A sample from the top of the unit (2.25m) was radiocarbon-dated to 39450±3263 yr B.P., suggesting this shallow phase occurred between ca 39,600-39,450 yr B.P.

The overlying sediment (2.25-6.60m) is grey silty clay. The change in lithology suggests increased water depth. Some aquatic pollen (including *Sparganium, Potamogeton, Myriophyllum* and *Spirogyra*: Huang et al., 1989) is present in this unit. These species are all characteristic of open water (rather than marsh conditions) and their presence is therefore consistent with increased water depth. The ostracode assemblage is characterised by abundant *Leucocythere postilirata*, suggesting fresh to semi-fresh conditions. Again, the change in the ostracode assemblage is consistent with deepening. By interpolation of the sedimentation rate (0.2434 cm/yr) between the radiocarbon date at the base of the unit and an overlying date from 11.6m, this unit is dated to between 39,450-37,670 yr B.P.

The overlying sediment (6.60-7.0m) is a second "grass peat", suggesting a return to shallow water and marshy conditions. The absence of aquatic pollen is consistent with this interpretation. The ostracode assemblage in the peat is characterised by *Limnocythere dubiosa*, consistent with increased salinity and shallowing. This unit is dated between ca 37,670-37,500 yr B.P.

The overlying sediment (7.0-10.0m) is grey and pale-green silty clay, suggesting increased water depth. The unit contains ca 2% aquatics, consistent with deepening. The ostracode assemblage is dominated by *Leucocythere postilirata*, again consistent with increased water depth. This unit is dated to between 37,500-36,270 yr B.P.

The overlying sediment (10.0-11.6m) is a third "grass peat", suggesting a return to shallower water conditions. The absence of aquatic pollen is consistent with this interpretation. However, the ostracode assemblage is still dominated by *Leucocythere postilirata* and thus does not support the idea of shallowing. A sample at 11.6m was radiocarbon-dated to 35,612 yr B.P., suggesting this peat phase occurred between 36,270-35,610 yr B.P.

The overlying sediment (11.6-12.4m) is grey and pale-green silty clay, suggesting increased water depth. The presence of ca 2% aquatics is consistent with increased water depth. The ostracode assemblage is dominated by *Leucocythere subculpta* with *Ilyocypris* spp. and *Candona* spp. (two species characteristic of fresh water), suggesting freshwater conditions. A sample at 12.4m was radiocarbon-dated to 29,441 yr B.P., suggesting this unit was formed between 35,610-29,440 yr B.P.

The overlying sediment (12.4-14.0m) is a fourth "grass peat", indicating decreased water depth after ca 29,440 yr B.P. The absence of aquatic pollen is consistent with this interpretation. The ostracode assemblage is dominated by *Leucocythere dubiosa*, consistent with shallow conditions.

The overlying sediment (14.0-19.2m) is grey silt with fine sand, suggesting increased water depth. The presence of sand suggests that the increase in water depth was less than indicated by the underlying lacustrine units. The presence of 5-6% aquatics in this unit is consistent with increased water depth. The ostracode assemblage is dominated by *Leucocythere postilirata, L. tropis*, consistent with increased water depth, but *Candona* spp. and *Ilyocypris* spp. disappears, suggesting that conditions were more saline than in the underlying lacustrine units. A sample at 12.4m was radiocarbon-dated to 29,441±884 yr B.P., suggesting this unit was formed between 27,860-22,740 yr B.P.

The uppermost sediment (19.2-19.5m) is a fifth "grass peat", indicating shallower water and marshy conditions. The absence of aquatic pollen is consistent with this interpretation. Ostracodes are not abundant, but the specimens present are mostly *Luecocythere dubiosa*, consistent with shallow-water conditions. This unit is dated to between 22,740-22,440 yr B.P. by extrapolation of the sedimentation rate (0.1014 cm/yr) between the radiocarbon dates at 11.6 and 12.4m.

Tagutuqiong Profile II is characterised by an alternation between lacustrine grey and white silt and silty clay deposits and "grass peats" (Li BY et al., 1991). Unfortunately, detailed descriptions of the deposits are not available and the depths of the "grass peat" units are unknown. It is therefore difficult to correlate this sequence with the sequence from Profile I. There are four thin "grass peat" layers in Profile II. The lowermost of these units was radiocarbon dated to 40,602±3320 yr B.P.; it is possible, given the large error bar on the date, that this unit correlates with the basal "grass peat" in Profile I (dated to between 39,450 and 39600 yr B.P.). Two dates of 36,454±847 and 30,302±685 yr B.P. were obtained from the middle of the second "grass peat" layer in Profile II. The oldest of these dates would suggest that the unit is correlated with the 3rd "grass peat" layer in Profile I (dated to 35610-36270 yr B.P.). However, the younger date suggests the unit should be correlated to the 4th "grass peat" layer in Profile I (dated to 27,860-29,440 yr B.P.). In either case, it is clear that part of the sequence in Profile I is missing from (or was not described in) Profile II. The top of the third "grass peat" unit in Profile II was dated to 25,560±674 yr B.P. Again, it is difficult to correlate this unit with the "grass peat" units in Profile I. The top of the uppermost "grass peat" layer in Profile II was dated to 18,187±167 yr B.P. It is plausible to assume that the uppermost two units in Profile II, which consist of lacustrine silty clays overlain by the uppermost "grass peat" unit, represent a return to deeper conditions postdating the deposition of the uppermost "grass peat" in Profile I (i.e. after ca 22,400 yr B.P.) and a final phase of shallower conditions which persisted until at least 18,100 yr B.P. However, without knowing the depths of the radiocarbon-dating samples from Profile II, it is not possible to determine how long each of these phases lasted.

Several former constructional shorelines, at 3m, 15m, 20m, 38m, 40m and 55m above the modern lake level respectively, can be clearly distinguished to the east of the modern lake (Li BY et al., 1991). These shorelines have not been studied or dated. However, deltaic sediments (Magazangbu Delta) occur in association with the 15m shoreline. This delta must therefore have been formed when the lake was ca 15m higher than today. A sample from the top of Magazangbu Delta (ca +15m above lake) was radiocarbon-dated to 6750±235 yr B.P. and a sample from the toe of Magazangbu Delta (height not given, but probably similar to that of the modern lake level) was radiocarbon-dated to 3330±200 yr B.P. This suggests that the Bangongcuo Lake was at least 15m higher than today from ca 6750 to 3330 yr B.P. (Li BY et al., 1991).

A 12.39m-long core (Bangongcuo Core), taken in 5m water depth from the southeastern part of the Bangongcuo (Li et al., 1994; Huang et al., 1996), provides a sedimentary record for the interval between 16,200 yr B.P. and the present (Li et al., 1994; Huang et al., 1996). Changes in water depth and water salinity are reconstructed from changes in lithology, ostracode assemblages and aquatic pollen in the Bangongcuo Core (Li et al., 1994; Huang et al., 1996). The chronology is based on 2 radiocarbon-dates from the core.

The basal sediment in Bangongcuo core (12.39-11.9m) consists of thin layers of silty clay, clay, sand and peat, indicating primarily shallow conditions but with short-lived oscillations in depth. There are relatively few ostracode remains but the assemblage is dominated by *Lymnocythere dubiosa*, consistent with relatively shallow conditions. A sample at 12.3m was radiocarbon-dated to 16,100±220 yr B.P. By interpolation of the sedimentation rate (0.129 cm/yr) between this date and a date from 3.35m, this unit was formed between ca 16,200-15,790 yr B.P.

The overlying unit (11.9-9.9m) is lacustrine grey and pale-green clayey silt, suggesting moderately deep water conditions. The presence of 5% *Myriophyllum* in aquatic pollen assemblage is consistent moderately deep water. The ostracode assemblage is dominated by *Candona gyirongensis* (typical of cold temperature < 10°C, fresh and deep water), with some *Leucocythere postilirata* and *L.* cf *subculpta* (typical of relatively fresh water and a wide range of shallow and muddy bottom conditions), suggesting fresh conditions. This unit was formed between 15,790 and 14,240 yr B.P.

The overlying sediment (9.9-9.8m) is lacustrine grey clay, suggesting increased water depth. The unit is devoid of aquatic pollen; the disappearance of *Myriophyllum* may indicate that the water depth became too great even for floating-leaved aquatics and therefore supports the lithological evidence suggesting the lake became deeper. There is no change in the ostracode assemblage, which is still dominated by *Candona gyirongensis* with some *Leucocythere postilirata* and *L.* cf *subculpta*, consistent with deep and fresh water. This unit formed between 14,240 and 13,620 yr B.P.

The overlying sediment (9.8-5.9m) is grey silt with molluscs shells and shell fragments. The change in lithology and the presence of shells suggest decreased water depth. The reappearance of *Myriophyllum* (5%) is consistent with shallowing. However, the ostracode assemblage is still dominated by *Candona gyirongensis* with *Ilyocypris biplicata, Leucocythere postilirata* and *L.* cf *subculpta*. The unit is dated to between 13,620-10,830 yr B.P.

The overlying sediment (5.9-5.12m) is silt containing at least two thin layers of fine sand. The presence of sand layers suggests a further decrease in water depth. The abundance of *Myriophyllum* decreased to <2%. Given that *Myriophyllum* is a floating-leaved aquatic, it is plausible that it should decrease in abundance as water depth decreased. The ostracode assemblage is dominated by *Limnocytherellina bispinosa* (a species tolerant of wide range of salinities up to 172.95 g/L: Li YF et al., 1991) with *Candona gyirongensis.* The abundance of *Limnocytherellina bispinosa* could be interpreted as indicating increased salinity and therefore shallower conditions than formerly. The unit is dated to between 10,830-10,220 yr B.P.

The overlying sediment (5.12-3.79m) is grey-black clay. The change in lithology indicates increased water depth. The abundance of *Myriophyllum* remains low, suggesting that the increase in water depth was not large. The abundance of ostracodes decreased. However, the presence of *Ilyocypris gibba, Ilyocypris cylindrituberosa* (both species tolerant of warm water temperature 10-20°C and a salinity of <1.78 g/L) and the presence of *Candona gyirongensis* is consistent with fresh water conditions. The unit is dated to between 10,220-9490 yr B.P.

The overlying sediment (3.79-2.4m) is silt with molluscs shells and shell fragments, suggesting decreased water depth. The ostracode assemblage is dominated by *Limnocytherellina bispinosa* and *L. trispinosa* occurs in the upper part of the unit. This marked change in the ostracode assemblage suggests increased salinity, and is consistent with decreased water depth. A sample at 3.35m was radiocarbon-dated to 9150±150 yr B.P., suggesting this unit was formed between 9490-6560 yr B.P.

The overlying sediment (2.4-2.3m) is fine sand, suggesting shallow water after 6560 yr B.P. The ostracode assemblage is still dominated by *Limnocytherellina bispinosa* with *L. trispinosa,* suggesting saline conditions.

The overlying sediment (2.3-2.0m) is clayey silt, suggesting a return to deeper water. The increased abundance of *Myriophyllum* (to 3%) is consistent with increased water depth. Although the ostracode assemblage is still dominated by *Limnocytherellina bispinosa* and *L. trispinosa,* the increased abundance of *Candona gyirongensis* and the appearance of *Candona candida* is consistent with decreased salinity. The unit is dated to between 6280-5460 yr B.P. by interpolation of the sedimentation rate (0.03661 cm/yr) between the radiocarbon date at 3.35m and the top of the core, which is assumed to be modern.

The overlying sediment (2.0-1.9m) is peat, suggesting decreased water depth after ca 5460 yr B.P. The absence of *Myriophyllum* is consistent with this interpretation. The ostracode assemblage is dominated by *Limnocytherellina bispinosa* and *Limnocythere dubiosa* (species tolerant of < 34 g/L) while the abundance of *Candona candida* and *Candona gyirongensis* decreased sharply*,* consistent with increased salinity and shallower conditions. This unit formed between ca 5460 and 5200 yr B.P.

The overlying sediment (1.9-1.1m) is clayey silt, suggesting increased water depth between 5200-3000 yr B.P. The presence of 2-3% *Myriophyllum* in aquatic pollen assemblage is consistent with increased water depth. The ostracode assemblage is dominated by *Limnocytherellina bispinosa* and *Limnocythere dubiosa*. However, the gradual decrease in the abundance of *Limnocytherellina trispinosa* and the increased abundance of *Candona gyirongensis, C. candida, Leucocythere postilirata* and *L. subculpta*, suggest that salinity decreased and is consistent with increased water depth.

The overlying sediment (1.1-1.0m) is peat, suggesting decreased water depth after ca 3000 yr B.P. However, *Myriophyllum* is still present (2-3%) in this unit. The ostracode assemblage is dominated by *Limnocytherellina bispinosa* with *Candona gyirongensis.*

The overlying sediment (1.0-0.5m) is silt with discrete layers of fine sand. The change in lithology suggests increased water depth, although the presence of sand layers indicates that the lake was still relatively shallow. The ostracode assemblage is marked by a further increase in the abundance of *Limnocytherellina bispinosa*, and the presence of *Limnocythere dubiosa* and *Candona gyirongensis.* The presence of *Limnocythere dubiosa* suggests the water was relatively saline, and is consistent with the idea that the increase in water depth was not large. This unit was formed between 2730 and 1370 yr B.P.

The overlying sediment (0.5-0.3m) consists of the remains of aquatic plants, mostly *Potamogeton* sp., suggesting shallow water conditions between ca 1370-540 yr B.P. The ostracode assemblage is still dominated by *Limnocytherellina bispinosa*. However, *Limnocythere dubiosa* disappears and *Candona gyirongensis* decreased in abundance. It is unclear how these changes in the ostracode record should be interpreted.

The uppermost sediment (0.3-0m) is fine sand, suggesting further decreased water depth after ca 540 yr B.P. The ostracode assemblage is still dominated by *Limnocytherellina bispinosa*. There is a marked decline in the abundance of *Candona gyirongensis*, consistent with decreased water depth.

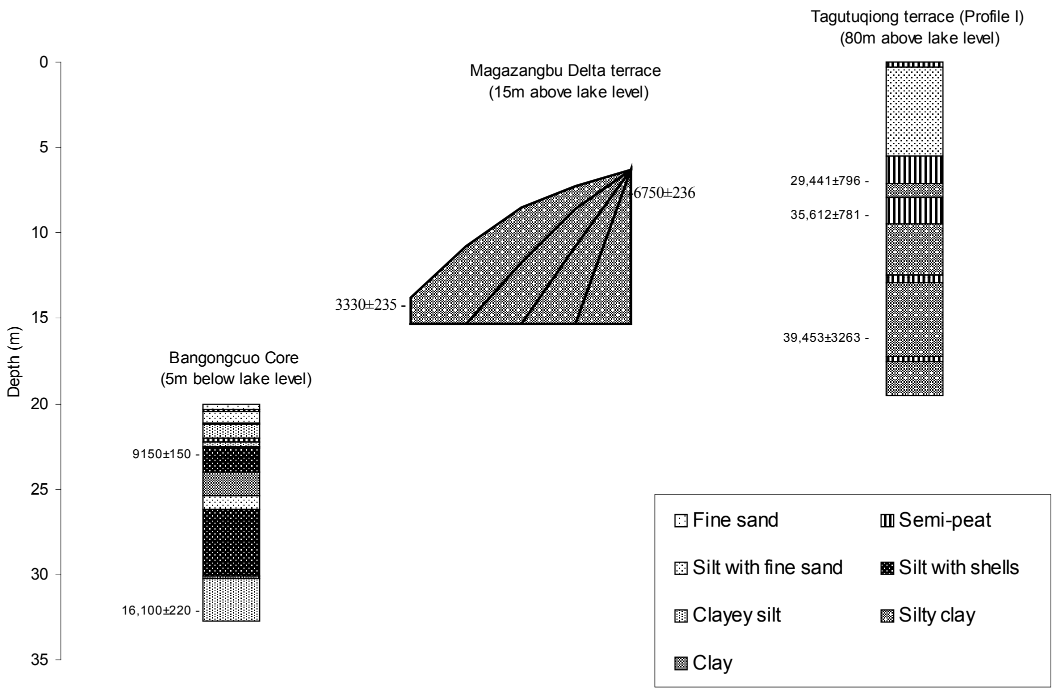
The sediments in the upper part of the Bangongcuo core were deposited synchronously with the formation of the Magazangbu Delta (between 6750 and 3330 yr B.P.). It is likely that the clayey silt deposits found between 2.3-2.0m (dated to 6280-5460 yr B.P.) and 1.9-1.1m (5200-3000 yr B.P.) were formed during the deep-water phase associated with the 15m shoreline and the formation of the delta. The core record makes it clear that lake levels were not consistently high during the interval 6750-3330 yr B.P., and thus suggest that the delta itself may have been formed during successive high water phases. However, if this correlation between the core deposits and the deltaic deposits is correct, it is possible to link the various records from the Bangongcuo basin and derive a relative coding scheme.

In the lake status coding, extremely low (1) is indicated by sand deposits in the Bangongcuo core; very low (2) by peat deposits or layers of aquatic plant remains in the Bangongcuo core; low (3) by silt with discrete sand layers and an ostracode assemblage dominated by *Limnocythere* spp. in the Bangongcuo core; moderately low (4) by silt deposits containing abundant mollusc shells in the Bangongcuo core; low-intermediate (5) by clayey silt deposits and an ostracode assemblage dominated by *Limnocytherellina* spp. in the Bangongcuo core and by formation of the the Magazangbu Delta; intermediate (6) by silts and an ostracode assemblage dominated by *Candona* spp. in the Bangongcuo core; moderately high (7) by lacustrine clays and an ostracode assemblage dominated by *Candona* spp. or *Ilyocypris* spp. in the Bangongcuo core; high (8) by "grass peat" deposits from the 80m terrace; very high (9) by silty sand deposits from the 80m terrace; and extremely high (10) by silty clay from the 80m terrace.

Radiocarbon dates

40,602±3320	plant remains, bottom of grass peat unit 1, Taketuqiong Profile II
39,453±3263	2.25m, plant remains, Taketuqiong Profile I
36,454±847	plant remains, middle of grass peat unit 2, Taketuqiong Profile II
35,612+865/-781	11.6m, plant remains, Taketuqiong Profile I
30,302±685	plant remains, middle of grass peat unit 2, Taketuqiong Profile II
29,441+884/-796	12.4m, plant remains, Taketuqiong Profile I
25,560±674	plant remains, top of grass peat unit 3, Taketuqiong Profile II
18,187±167	plant remains, top of grass peat unit 4, Taketuqiong Profile II
16,100±220	12.3m, clay, Bangongcuo Core
9150±150	3.35m, silt, Bangongcuo Core
6750±235	ca +15m above lake, the top of 15m-high lake-delta terrace, Magazangbu
3330±200	the front of 15m-high lake-delta terrace, Magazangbu

The samples were dated in ^{14}C Lab, Geography Institute, Chinese Academy of Science.



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Coding

pre 39,600 yr B.P.	extremely high (10)
39,600-39,450 yr B.P.	high (8)
39,450-37,670 yr B.P.	extremely high (10)
37,670-37,500 yr B.P.	high (8)
37,500-36,270 yr B.P.	extremely high (10)
36,270-35,610 yr B.P.	high (8)
35,610-29,440 yr B.P.	extremely high (10)
29,440-27,860 yr B.P.	high (8)
27,860-22,740 yr B.P.	very high (9)
22,740-22,440 yr B.P.	high (8)
22,400-18,400 yr B.P.	extremely high (10)
19,200-18,180 yr B.P.	high (8)
18,180-16,200 yr B.P.	uncoded
16,200-15,790 yr B.P.	low (3)
15,790-14,240 yr B.P.	intermediate (6)
14,240-13,620 yr B.P.	moderately high (7)
13,620-10,830 yr B.P.	moderately low (4)
10,830-10,220 yr B.P.	low (3)
10,220-9490 yr B.P.	moderately high (7)
9490-6560 yr B.P.	moderately low (4)
6560-6280 yr B.P.	extremely low (1)
6280-5460 yr B.P.	low-intermediate (5)
5460-5200 yr B.P.	very low (2)
5200-3000 yr B.P.	low-intermediate (5)
3000-2730 yr B.P.	very low (2)
2730-1370 yr B.P.	low (3)
1370-540 yr B.P.	very low (2)
540-0 yr B.P.	extremely low (1)

Preliminary coding: 08-02-1999
Final coding: 23-01-2001
Coded by GY and SPH

3.34. Cuona Lake (CoNag Lake) , Xizang (Tibet) Autonomous Region

Cuona Lake (called CoNag by Shen and Xu, 1994) (32.03°N, 91.47°E, 4590m above sea level) is a large freshwater (0.14g/L) lake in the Ando Region of the inner part of the Tibetan Plateau (Shen and Xu, 1994). The modern lake has an area of 174 km^2 and a water depth of >10m. The catchment area is ca 10-15 times the lake area. The basin is surrounded by mountains which reach elevations of 4900-5300 m a.s.l. The lake is fed by direct precipitation and river discharge. Three large rivers drain from the high mountains around the basin. These rivers are fed by precipitation, glacial meltwater and snowmelt from the mountains (Shen and Xu, 1994). Although there are glaciers within the lake catchment, their contribution to the overall water budget is thought to be small. There is an outflow from the lake which ultimately discharges into the Nu Jiang-Salween River, which finally reaches the Indian Ocean. However, increasing aridity in recent decades has resulted in lowering of the lake level and reduced discharge, so it is thought that the lake will become closed within a relatively few years. The Cuona basin originated by faulting during the early Tertiary. Although tectonic activity and uplift have continued during the Quaternary, the impact of tectonism on the Cuona basin during the last ca 35,000 years is negligible. The bedrock is Jurassic, red, muddy sandstone and occurs at depth of ca 12-15m beneath the lake bottom (Shen and Xu, 1994). Subaerial sediments derived from weathering of the Jurassic bedrock have a characteristic red colouration and this colour is also characteristic of modern shallow water lacustrine deposits (Shen and Xu, 1994). The climate in the Ando region is cold (-3°C mean annual temperature) and dry (411.6mm total annual precipitation, 1770mm annual evaporation) (Shen and Xu, 1994). The Ando Region is characterised by alpine steppe and tundra vegetation, both dominated by *Stipa* spp.

There appear to be four lacustrine terraces to the west of the modern lake, although only the two highest terraces (T3, T2) have been described. The top of the highest terrace (T3), which occurs as discontinuous fragments, is at 130m above modern lake level (i.e. at 4720 m a.s.l.). The top of the second terrace (T2) occurs at 20m above modern lake level (i.e. at 4610 m a.s.l.). It is not clear whether the T3 terrace is an erosional or a depositional feature. However, a core from the T2 terrace shows that it consists of lake bottom sediments. The T2 terrace is therefore erosional in origin. The T3 terrace is undated. Deposits from the T2 terrace have been radiocarbon dated, and the record appears to begin at ca 35,000 yr B.P. Given their relative position, the highest (T3) terrace must be at least this old although it could be considerably older. Lacustrine terraces of a similar elevation (i.e. 4720 m a.s.l.) to the T3 terrace have been found in the Zigetangcuo basin, which lies ca 100km west of Cuona. On the basis of this apparent congruity of elevation, Shen and Xu (1994) claim that the two lake basins were joined to form a single mega-lake sometime before 35,000 yr B.P. However, given the distance between the two basins, the absence of evidence of a significant spillway through the divide between the two basins, and the fact that there is no direct evidence to show that the two sets of terraces were joined, it seems unlikely that Zigetangcuo and Cuona were part of a single lacustrine system in the Late Quaternary.

The T2 terrace at Cuona lies ca 30m from the edge of the modern lake. An 8m-long (Core SG-89-2) provides a sedimentary record from this terrace (Shen and Xu, 1993; Shen and Xu, 1994). Shen and Xu (1994) have reconstructed changes in water depth on the basis of changes in the lithology and aquatic macrofossil assemblages from the core sediments. They claim that the core record covers the interval from ca 35,000 to ca

12,000 yr B.P. However, there are only four radiocarbon dates from the core, the youngest of which is 20,916±1205 yr B.P., so it is possible that all of the sediments are of glacial age.

The basal sediments (8.0-8.2m) in Core SG-89-2 are white quartz sand and gravel. Characteristic markings on the surface of the quartz grains, such as en enchelon concave fractures, indicate that this material was produced by glacial erosion. This unit is therefore interpreted as a glacial deposit formed before ca 35,000 yr B.P. (Shen and Xu, 1994). During the deposition of this unit, there was no lake in the Cuona basin.

The overlying sediment (8.0-7.5m) is black clay. There are thin bands of microfossils within the clay. The fine-grained nature of the sediment, and the preservation of thin laminar structures within the unit is consistent with deposition in a relatively deepwater lake. The colour of the deposits is also indicative of deepwater conditions, since shallower lacustrine deposits tend to be red in colour. A sample from the lower boundary of this lacustrine unit is radiocarbon-dated to 35,000 yr B.P. and a sample at 7.6m to 25,397±964 yr B.P. Thus, a deepwater lake was present in the Cuona basin between 35,000-25,000 yr B.P. (Shen and Xu, 1994). The colour of the lacustrine deposits is not consistent with this deepwater lake being fed by glacial meltwater, and the abrupt nature of the contact between the lacustrine clay and the underlying glacial deposits is consistent with a sedimentary hiatus between the two units. This suggests that the glaciers had retreated from the Cuona basin considerably before the onset of lacustrine conditions.

The overlying sediments in Core SG-89-2 are coarse (7.5-7.3m) and fine (7.3-7.1m) grey-coloured sands. These sands were probably formed as a beach deposit. The change in lithology therefore indicates a considerable decrease in water depth. Interpolation of the sedimentation rate (0.0272 cm/yr) between the radiocarbon date from the top of the underlying lacustrine clay and a date of 21,247±1130 yr B.P. from an overlying unit suggests that these sand units were formed between 25,000-23,600 yr B.P.

The overlying sediments (7.1-6.9m) are greyey-black silty clay. The relatively fine-grained nature of the sediments indicates increased water depth after ca 23,600 yr B.P. However, both the colour and the presence of silt indicate that the lake was not as deep as during the first deepwater lacustrine phase.

A return to shallow-water conditions is characterized deposition of a layer of fine, red-coloured, beach sands (6.6-6.9m). The red-colouration suggests that this material was derived by reworking of nearshore deposits that were themselves ultimately derived from weathering (and reworking) of bedrock from the catchment. This beach sand unit is dated, by interpolation of the sedimentation rate between the bracketing radiocarbon dates, to ca 22,800-21,500 yr B.P.

The overlying sediment is lacustrine reddish sandy clay (4.4-6.6m). The change in lithology is consistent with an increase in water depth. The unit probably represents deposition in the nearshore (littoral) zone. The red colouration is consistent with this interpretation. The presence of four discrete layers of fresh-water mollusc (spiral) shells (6.5-6.6m, 6.1-6.2m, 5.7-5.8m, 4.8-5.9m and 4.4-4.6m) within the unit is also consistent with shallow-water, nearshore deposition. A sediment sample from 6.5m is radiocarbon-dated to 21,347±1130 yr B.P. and a sample from 4.4m dated to 20,916±1205 yr B.P., indicating that this phase of shallow-water deposition occurred between 21,500-20,900 yr B.P. These two samples represent the youngest radiocarbon dates from the core, and all the overlying units are dated by extrapolation of the sedimentation rate (0.0387cm/yr) from between them.

The overlying sediment (4.4-4.25m) is grey-black clay. The change in lithology and in colour both indicate increased water depth. This unit is dated to 20,900-20,500 yr B.P.

The overlying sediment (4.25-3.8m) is reddish silty clay. The coarser nature of the sediments and the red colour indicate that the water became shallower. Extrapolation of the sedimentation rate from the underlying radiocarbon dates indicates that this unit was formed between 20,500-19,400 yr B.P.

The overlying sediment (3.8-3.5m) is grey-black clay. The change in lithology and colour indicate a return to deepwater conditions. The age of these deposits is estimated to between 19,400-18,600 yr B.P.

The overlying sediment (3.4-3.25m) is reddish silty clay. The coarser nature of the sediments and the red colour suggest decreased water depth. The unit is estimated to have formed between 18,600-18,300 yr B.P.

The overlying sediment (3.4-3.25m) is grey-black clay. Again, the change to fine-grained sediments and the change in colour are both consistent with increased water depth. By extrapolation, the unit formed between 18,300-17,900 yr B.P.

The overlying unit (3.25-3.2m) is reddish fine sand. This unit is interpreted as a beach deposit, and marks a significant shallowing of the lake after ca 17,900 yr B.P.

The overlying grey-black clay (3.2-2.8m) indicates a return to deeper water conditions between ca 17,800-16,800 yr B.P.

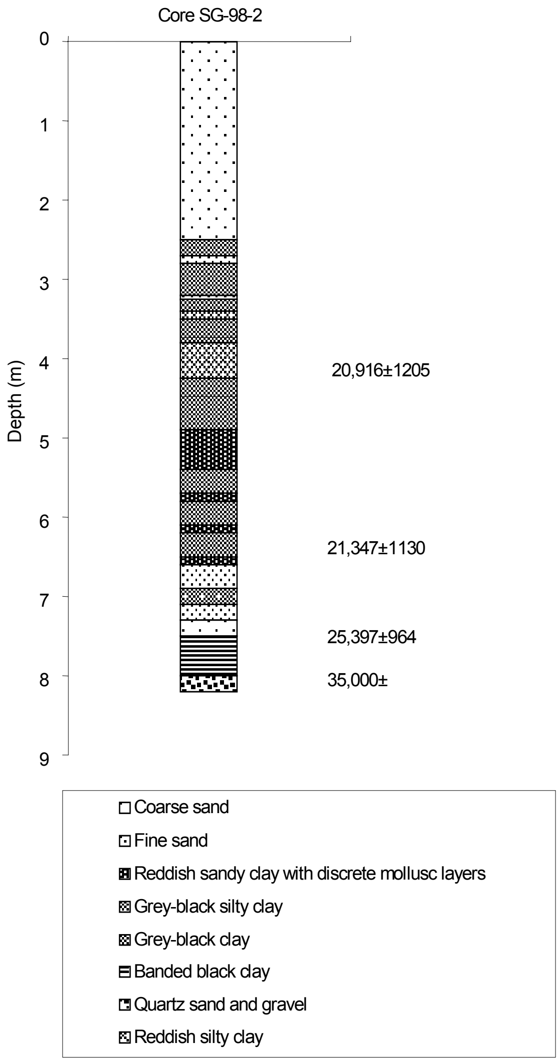
The overlying unit (2.8-2.5m) is reddish, fine sand, assumed to be a beach deposit and therefore to indicate shallow water conditions after ca 16,800 yr B.P.

The overlying grey-black clay (2.7-2.5m) indicates increased water depth between ca 16,500-16,000 yr B.P.

The uppermost sediment (2.5-0m) is reddish coarse sand characteristic of beach deposits, suggesting a return to shallower conditions after 16,000 yr B.P. Shen and Xu (1993, 1994) estimated that this unit was formed between ca 16,000-12,000 yr B.P. by extrapolation of the sedimentation rate between the underlying radiocarbon dates. However, beach sands can be deposited relatively fast, and there is no reason to suppose that this unit represents ca 4000 years of deposition. We therefore prefer not to ascribe an upper age limit to the unit.

The erosional nature of the T2 terrace suggests that there was a substantial and perhaps abrupt lowering of lake level sometime after the deposition of the uppermost coarse sand unit. Given the inferred presence of two lower terraces, this lowering may have taken place in a number of stages each of which was marked by erosion and removal of lake bottom material from the central part of the palaeolake. However, there are no descriptions of the material within the lower terraces or of the deposits underlying the modern lake. It is therefore not possible to reconstruct a history of lake changes during the late glacial and Holocene. However, the modern lake is apparently at the lowest level (4590m a.s.l) recorded in the history of Cuano Lake (Shen and Xu, 1994).

In the status coding, very low (1) is indicated by the modern lake level of 4590 m a.s.l.; low (2) by coarse or fine beach sands in the T2 terrace core; moderately low (3) by red sandy clays containing freshwater molluscs and assumed to be deposited in the littoral (nearshore) zone; intermediate (4) by red silty clay nearshore deposits; moderately high (5) by grey-black lacustrine silty clay; high (6) by grey-black lacustrine clay; and very high (7) by finely-banded, black lacustrine clay.



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Radiocarbon dates

35,000±?	8.2m, black clay, Core SG-89-2	Date from Kiel University
25,397±964	7.6m, black clay, Core SG-89-2	Date from Lanzhou ^{14}C lab
21,347±1130	6.5m, sandy clay, Core SG-89-2	Date from Lanzhou ^{14}C lab
20,916±1205	4.4m, sandy clay, Core SG-89-2	Date from Lanzhou ^{14}C lab

Coding

ca 35,000-25,000 yr B.P.	very high (7)
25,000-23,600 yr B.P.	low (2)
23,600-22,800 yr B.P.	moderately high (5)
22,800-21,500 yr B.P.	low (2)
21,500-20,900 yr B.P.	moderately low (3)
20,900-20,500 yr B.P.	high (6)
20,500-19,400 yr B.P.	intermediate (4)
19,400-18,600 yr B.P.	high (6)
18,600-18,300 yr B.P.	intermediate (4)
18,300-17,900 yr B.P.	high (6)
17,900-17,800 yr B.P.	low (2)
17,800-16,800 yr B.P.	high (6)
16,800-16,500 yr B.P.	low (2)
16,500-16,000 yr B.P.	high (6)
16,000-ca 15,500 yr B.P.	low (2)
15,500-0 yr B.P.	not coded
0 yr B.P.	very low (1)

Preliminary coding: 19-11-1998
Final coding: 26-07-2000
Coded by GY and SPH

3.35. Hongshanhu Lake, Xizang (Tibet) Autonomous Region

Hongshanhu Lake (37.45°N, 78.99°E, 4870m above sea level) is a shallow, closed salt lake in the West Kunlun Mountains on the western Tibetan Plateau. The basin originated by faulting, and the bedrock in the basin is Jurassic-Cretaceous sandstone (Wang et al., 1987). The lake has an area of 4.3km^2. A few brooks, draining from the high mountains, enter the lake (Li et al., 1995). The climate in the region is extremely cold and dry, with mean annual temperature 0° C, and total annual precipitation of 50mm but annual evaporation of 2500mm (Li et al., 1994).

A lake terrace to the southeast of the lake provides evidence of a former high stand of the lake. The terrace deposits have not been investigated in detail. However, a 6m-high natural section has been cut into the lowermost part of the terrace. The top of the natural section is at ca 4876 m a.s.l. (+6m above modern lake level). The exposed sediments provide a record covering the period between 17,000-13,000 yr B.P. (Li et al., 1994). Changes in water depth and salinity are reconstructed on the basis of changes in lithology, ostracode assemblages and aquatic macrofossils, and follow the interpretations of Li et al. (1995). The chronology is based on four radiocarbon dates on bulk sediments from the profile.

The basal sediment (below 5.6m) is pale-black lacustrine silty clay. The unit contains aquatic plant remains (specific taxa are not identified by Li et al., 1994). The ostracode assemblage is dominated by *Candona candida* and *C. neglecta* (deep- and fresh-water species: Huang et al., 1985) and *Leucocythere mirabilis* (tolerant of a wide range of conditions, from fresh to highly saline water up to 150 L/g: Li et al., 1994) with *Ilyocypris gibba* (a fresh-water tolerant species: Li et al., 1994), and probably indicates fresh water conditions (Li et al., 1994). The lithology and ostracode assemblage are consistent with moderately deep water. A sample from near the upper boundary of this unit (5.7m) was radiocarbon-dated to 17,015±151 yr B.P.

The overlying sediment (5.6-4.8m) is grey laminated silty clay, with aquatic plant remains. The preservation of laminations suggests increased water depth. The ostracode assemblage is characterised by the dominance of *Candona candida* and *C. neglecta*, the continued presence of *Ilyocypris gibba*, and a decrease in *Leucocythere mirabilis*. This shift in dominance might reflect decreased water salinity, consistent with deepening. By interpolation of the sedimentation rate between the available radiocarbon dates, this unit was deposited before 16,800 yr B.P.

The overlying sediment (4.8-4.4m) is grey-pale white silt. Both the change in texture and the non-laminated nature of the sediments suggests decreased water depth. There are some aquatic plants present in this unit, consistent with decreased water depth. The ostracode assemblage is still dominated by *Candona candida* and *C. neglecta* with *Leucocythere mirabilis*, but *Ilyocypris gibba* disappears. The increased importance of *Leucocythere mirabilis* and the disappearance of *Ilyocypris gibba* may reflect a slight increase in salinity, consistent with shallower water.

The overlying sediment (4.4-2.2m) is grey, light-yellow and white laminated silty clay, suggesting a return to relatively deep water conditions. The disappearance of aquatic plants is consistent with an increase in water depth. The ostracode assemblage is dominated by *Candona candida* and *C. neglecta*, but *Leucocythere mirabilis* is not present. The disappearance of this species may indicate decreased salinity, consistent with an increase in water depth. Two samples from 4.1m and 2.2m were radiocarbon-dated to 16,428±132 yr B.P. and 15,310±178 yr B.P. respectively, suggesting this unit was formed between 16,500-15,300 yr B.P.

The overlying sediment (2.2-0.7m) is a grey silty clay, with thin layers of yellow silt. The presence of silt suggests shallow water. The unit contains abundant aquatic plant remains, consistent with decreased water depth. The ostracode assemblage is dominated by *Limnocythere dubiosa*, *Candona candida* and *C. neglecta.* However, *Cyprideis torosa* (a species tolerant of salinities up to 120 g/L) occurs in moderate abundance. The increased importance of *Limnocythere dubiosa* and the presence of *Cyprideis torosa* indicates increased salinity. This unit is dated to ca 15,310-13,750 yr B.P. by interpolation of sedJimentation rate (0.096 cm/yr) between radiocarbon dates.

The uppermost sediment (0.7-0m) is pale-yellow silty clay. The ostracode assemblage is dominated by *Limnocythere dubiosa* and *Candona* spp., but *Cyprideis torosa* becomes less abundant, indicating a decrease in salinity. A sample at the lower boundary (0.7m) of the unit was radiocarbon-dated to 13,750±120 yr B.P. and the upper boundary is estimated by extrapolation of the sedimentation rate (0.096 cm/yr) to ca 13,000 yr B.P.

The sedimentary record after 13,000 yr B.P., which must be preserved in the sediments in the upper part of the lacustrine terrace, has not been studied. The modern lake has the lowest lake level during its history (Li et al., 1995).

In the status coding, very low (1) is indicated by modern lake level; low (2) by silty-clay with abundant *Cyprideis torosa*; moderately low (3) by silty clay with sparse *Cyprideis torosa*; intermediate (4) by non-laminated silty clay or silt with a mixed ostracode assemblage including *Candona candida*, *C. neglecta*, *Leucocythere mirabilis* and *Ilyocypris gibba*; high (5) by laminated silty clay with *Candona candida*, *C. neglecta* and *Ilyocypris gibba*; and very high (6) by laminated silty clay with only *Candona* spp.

Radiocarbon dates

17,015±151	5.7m, silty clay
16,428±132	4.1m, silty clay
15,310±178	2.2m, silty clay
13,750±120	0.7m, silty clay (13,752±120 yr B.P. given in Li et al., 1995)

(The four samples were dated in Institute of Geography, Chinese Academy of Science)

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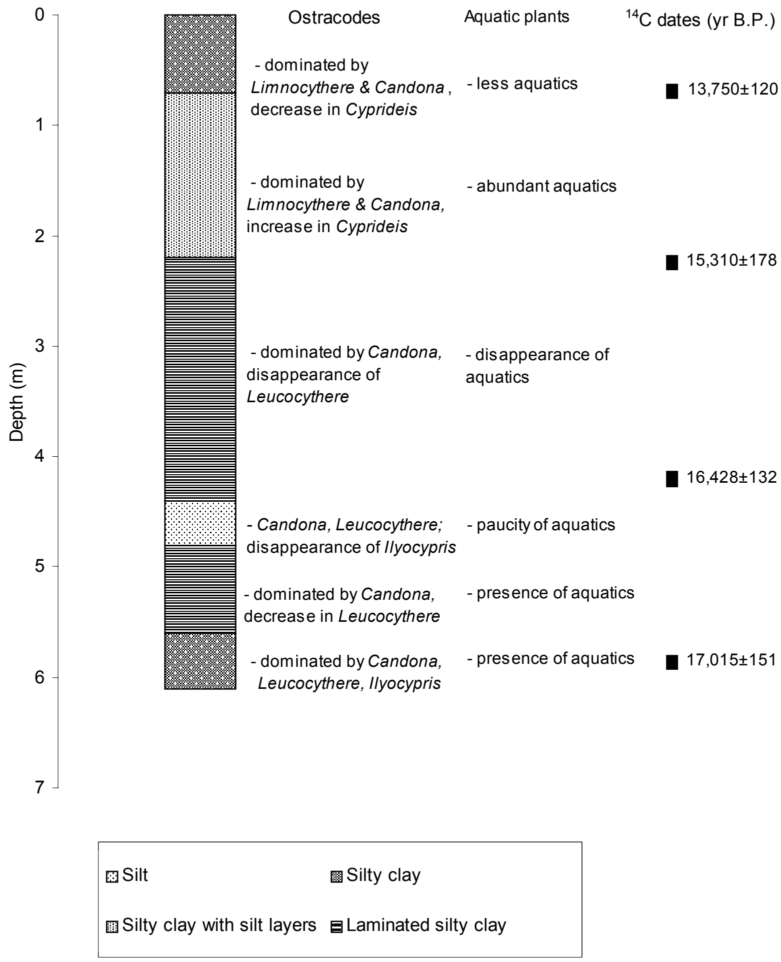


Coding

17,200-17,000 yr B.P.	intermediate (4)
17,000-16,800 yr B.P.	high (5)
16,800-16,500 yr B.P.	intermediate (4)
16,500-15,300 yr B.P.	very high (6)
15,300-13,700 yr B.P.	low (2)
13,700-13,000 yr B.P.	moderately low (3)
13,000-100 yr B.P.	not coded (no record)
0 yr B.P.	very low (1)

Preliminary coding: 24-11-1998
Second coding: 15-12-1998
Third coding: 4-1-1999
Final coding 27-3-1999
Coded by GY and SPH

3.36. North Tianshuihai Lake, Xizang (Tibet) Autonomous Region

North Tianshuihai (35.70°N, 79.37°E, 4797-4800m above sea level) is a salt lake in the West Kunlun Mountains, on the western Tibetan Plateau. The lake lies in a large structural basin, with an area of ca 18,800 km^2 and a threshold elevation of 4935 m a.s.l. (Wang et al., 1990). This structural basin contains a number of inset lacustrine basins, including those of the modern lakes Guozacuo, Akesaiqin, Tianshuihai, North Tianshuihai and Kushiuhai. The inset basins originated through faulting. Although tectonism is still active in this region, the effects of uplift on the lake system is thought to be small compared to the effects of climate change (Wang et al., 1990).

During the early Quaternary, the North Tianshuihai basin was part of a much larger palaeolake, which unified the modern lakes of Akesaiqin, Tianshuihai, North Tianshuihai and Kushiuhai (Wang et al., 1990; Li et al., 1991). The extent of the mega-palaeolake is defined by intermittent shorelines between 4880-4890m a.sl. and is estimated to have been ca 2650 km^2 (Wang et al., 1990) and the catchment area was equivalent to the whole of the structural basin (i.e. ca 18,800 km^2). However, Guozacuo lake does not appear to have been connected with the mega-palaeolake at any time. On the basis of a single radiocarbon date, with an age >45,000 yr B.P., Wang et al. (1990) suggest that the mega-palaeolake existed before 45,000 yr B.P. The mega-palaeolake was fresh and had an overflow to the Karakala River.

As climate became more arid after ca 45,000 yr B.P., lake level fell and the individual basins became distinct. Akesaiqin Lake was originally connected to Tianshuihai Lake via a river channel, but by the early Holocene this connection was broken. Tianshuihai (4830 m a.s.l.) and North Tianshuihai (4800m a.sl.) lakes are still part of the same system, since they both have an intermittent overflow to Kushuihai Lake (4754 m a.s.l.). However, there is no direct surface or subsurface connection between Tianshuihai and North Tianshuihai, and given that the distance from Tianshuihai to Kushuihai is >30 km and from North Tianshuihai to Kushuihai is ca 20 km, it seems likely that the history of the two lakes during the last 45,000 yr B.P. can be considered as independent. Akesaiqin Lake (4840 m a.sl.) lies 65km southeast of North Tianshuihai Lake, and ca 40km eastsoutheast of Tianshuihai. There is no surface or subsurface connection between this lake and the other lakes in the structural basin. Its history over the last 45,000 yr B.P. can therefore also be considered separately from that of the other lakes in the basin.

The modern North Tianshuihai Lake consists of five sub-basins connected by narrow channels, and has a total area of 8 km^2. The area of the largest of the sub-basins is 3.7 km^2 (Wang et al., 1990). The lake has a maximum depth of 12.6m (Li et al., 1991). Lake water is supplied by springs in the piedmont of the Kunlun Mountains. A seasonal outlet overflows westnorthwestward to Kuishuihai Lake, which is closed. According to Wang et al. (1990), this seasonal overflow is decreasing due to increasing aridity and it is likely that the North Tianshuihai, Tianshuihai and Kushuihai lakes will be completely separated from one another in the near future. Modern sediments in the North Tianshuihai Lake show a characteristic distribution, with laminated clays and silty clays in the deepest part of the lake grading to non-laminated silts, and silts and sands in the nearshore zone, and sand and gravels in the littoral zone (Li et al., 1991). The aquatic plant, *Potamogeton* spp. is found in the lake in water depths between 1-8m, and *Carex moorcrofrii* occurs along the lake shoreline and within the catchment. The modern regional climate is extremely cold and dry, with -5°C mean annual temperature and mean annual precipitation between 20-40 mm (Wang et al., 1990) and mean annual evaporation of ca 2500 mm.

Geomorphological investigations have identified a number of lacustrine terraces around North Tianshuihai Lake, which provide evidence of former high lake stands during the last ca 18,000 yr B.P. (Wang et al., 1990; Li et al., 1991). Although some of these terraces contain beach material, and could therefore be constructional terraces, the remaining terraces are composed of lacustrine material and must therefore have been created by erosion of more extensive deposits either during lake regressions from former high stands or during subsequent transgressions. Six terraces have been identified on the southern side of the lake, at 4850-4852m, 4840-4843m, 4831m, 4817m, 4809-4812m and 4805-4806 m a.s.l. On the basis of elevational correlation, four of these terraces have been recognized on the northwestern side of the lake at 4850-4855m, 4832-4835m, 4808-4810m and 4802-4806m. The reasons for the absence of the 4840-4843m and 4817m terraces to the northwest of the lake are unclear, although not unexpected if most of the terraces are erosional in origin. Profiles (T50, S10, T8, S9, T10, T5) through each of the four terraces on the northwest side of the lake provide detailed records of changes in lithology and aquatic pollen assemblages, which have been used to reconstruct changes in lake level through time (Wang et al., 1990; Li et al., 1991). The interpretations here basically follow those of the original authors. The chronology is based on 8 radiocarbon dates from the profiles.

Profile S10 provides a 7m-deep section through the 4832-4835m terrace northwest of the lake. The top of the profile at the profile location is 4835 m a.s.l. The basal sediment (6-7m) is laminated lacustrine clay, which directly overlies slate bedrock. The overlying unit (6-5m) is laminated silty clay with abundant aquatic remains. The increased silt content is consistent with a slight decrease in water depth. The uppermost unit (0-5m) is non-laminated lacustrine silt, indicating a further decrease in water depth. Samples from 6.5m and 3m were radiocarbon-dated to 17,700±174 and 17,480±155 yr B.P. respectively, suggesting that the deposits in this terrace were formed between ca 17,750 and 17,450 yr B.P. The water level must have been > 4830m when the laminated basal units were formed and, given that laminated deposits are unlikely to have formed in water depths of <10m, was probably ≥ 4840 m a.s.l. (i.e. at least +43m above modern lake level). The water depth during the formation of the upper units was ≥ 4835 m a.s.l. (i.e. +38m above modern lake level).

Profile S9 provides a 2.5m thick section through the 4802-4806m terrace northwest of the lake. The top of the terrace at this site is at 4802m. The basal sediment (2.5-1.7m) is lacustrine in origin. The overlying unit (1.6-1.7m) is sand, and has been interpreted as a beach deposit (Wang et al., 1990). The lake level must have been at ca 4800-4801 m a.s.l. when this unit formed. A sample from the overlying unit (1.50-1.55m) was radiocarbon-dated to 17,360±180 yr B.P., indicating that the beach unit was formed before ca 17,400 yr B.P. Thus, the record from Profile S9 indicates that lake level fell from ca ≥ 4835 m a.s.l. to ca 4801m between 17,450 and 17,400 yr B.P.

The overlying sediment in Profile S9 (1.2-1.6m) is silty clay, containing abundant aquatic plant remains, and characteristic of littoral or nearshore deposition. This change in lithology indicates increased water depth. Interpolation of the sedimentation rate (0.0296 cm/yr) between the date of 17,360±180 yr B.P. from near the base of this unit and the date of 15,670±110 yr B.P. from a depth of 1.0m in the profile, indicates that this phase of increased water depth persisted until ca 16,345 yr B.P.

The overlying sediment in Profile S9 (1.1-1.2m) is a beach sand, indicating that lake level fell somewhat after 16,345 yr B.P. However, a return to deeper water conditions is indicated by the deposition of lacustrine silty clay (0.9-1.1m). A sample from near the top of this unit (1.0m) was radiocarbon dated to 15,670±110 yr B.P., indicating that the return to deeper water conditions occurred between 16,000 and 15,595 yr B.P.

A third beach-sand unit (0.8-0.9m) indicates a return to somewhat shallower conditions. By interpolation of the sedimentation rate (0.1339 cm/yr) between the available radiocarbon dates, this interval is dated to 15,595-15,520 yr B.P. The beach deposit is overlain by lacustrine silty clay (0.6-0.8m) , indicating a return to slightly deeper-water conditions between 15,520-15,370 yr B.P.

The uppermost unit (0.6-0m) is black clay, containing abundant aquatic plant remains, mostly of *Potamogeton* spp. The change in lithology suggests an increase in water depth after ca 15,370 yr B.P. A sample from a depth of 0.25m was radiocarbon-dated to 15,110±30 yr B.P. Extrapolation of the sedimentation rates on the underlying units suggests that this deeper-water phase persisted until ca 15,300 yr B.P.

Profile T5 provides a 4m-deep section through the 4802-4806m terrace northwest of the lake. The elevation of the top of the profile is 4805m. The basal unit (2-4m) is lacustrine silty clay. A sample from ca 3m of 14,500±340 yr B.P. The deposition of this unit indicates shallower conditions than the *Potamogeton*-unit found at the top of S9. Using a sedimentation rate of 0.134 cm/yr, calculated from similar silty clay deposits in S9, the duration of this lower-water phase would be from ca 15,250 to 13,750 yr B.P. The dating of the bottom boundary is consistent with the dating of the upper boundary of the *Potamogeton*-rich layer in S9. Thus, the dating is consistent with the interpretation that the lake level fell after ca 15,300 yr B.P. A similar unit is found in the T10 profile between the elevation of 4809-4810m. However, since this material is undated, the T10 profile cannot be used to infer the possible maximum depth of the lake between 15,250 and 13,750 yr B.P.

The overlying unit in T5 (0-2m) is laminated silty clay and silt, with abundant green algae. The preservation of laminations indicates that the lake became deeper ca 13,750 yr B.P.

A 6-m deep profile (Profile T50) from the 4850-4855m terrace northwest of the lake, contains lacustrine grey-white sandy clay (1-6m) overlain by lake beach deposits (0-1m). The top of the profile at this location is at an elevation of 4850m. The lacustrine unit contains abundant plant remains, including *Potamogeton* spp., and likely represents a nearshore deposit. Thus the profile reflects a regressional phase at the time when the lake was rather deep. A sample from 1.5m was radiocarbon-dated to 13,070±200 yr B.P. Thus, the record shows the lake level must have been ≥ 4850m (i.e. +53m above modern lake level) before 13,000 yr B.P. and subsequently fell to 4850m.

There is no lithological record of the period after ca 13,000 until ca 11,300 yr B.P.

Profile T8 is a 4m-thick section from the 4808-4810m terrace northwest of the lake. The top of the terrace at this site is at 4808m. The sediments (0-4m) overlie bedrock, and consist of grey-white sandy silt with thin and ripple bedding characteristic of nearshore lacustrine deposition. On the assumption that deposits with these kinds of structures are deposited in water depths of ca 2m, these sediments indicate a lake level of ca 4810m (i.e. +13m above modern lake level). A sample from near the base of the profile (3.8-

4.0m) was radiocarbon-dated to 11,235±650 yr B.P., indicating this relative high stand occurred after ca 11,300 yr B.P.

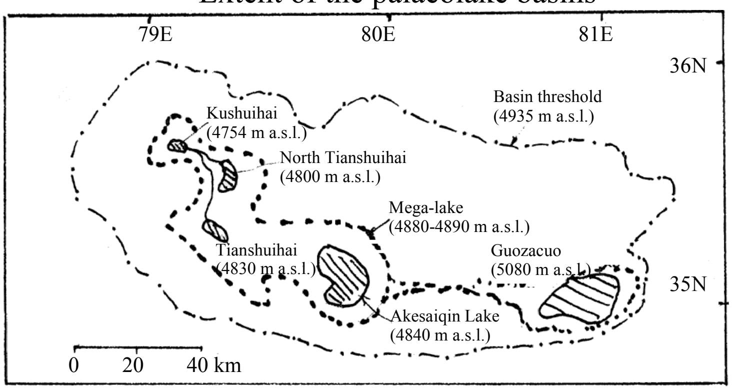
In the status coding, very low (1) is indicated by lake levels lower than 4800m; low (2) by lake levels between 4800-4802m; moderately low (3) by lake levels between 4802-4805m; intermediate (4) by lake levels between 4805-4810m; moderately high (5) by lake levels between 4810-4815m; high (6) by lake levels between 4815-4840m; very high (7) by lake levels between 4840-4850m; and extremely high (8) by lake levels higher than 4850m.

Radiocarbon dates

17,700±174	6.5m, clay, Profile S10 *
17,480±155	3.0m, silt, Profile S10 *
17,360±180	1.50-1.55m, clay, Profile S9 *
15,670±110	1.0m, silty clay, Profile S9 *
15,110±30	0.25m, aquatic plant, Profile S9 *
14,500±340	3m, silty clay, Profile T5
13,070±200	1.5m, sandy clay, Profile T50
11,235±650	3.8-4.0m, silt, Profile T8

(The 5 samples marked with * were dated in ^{14}C Lab, Department of Geography, Lanzhou University; the rest in ^{14}C Lab, Geo and Ocean Science, Nanjing University)

Extent of the palaeolake basins



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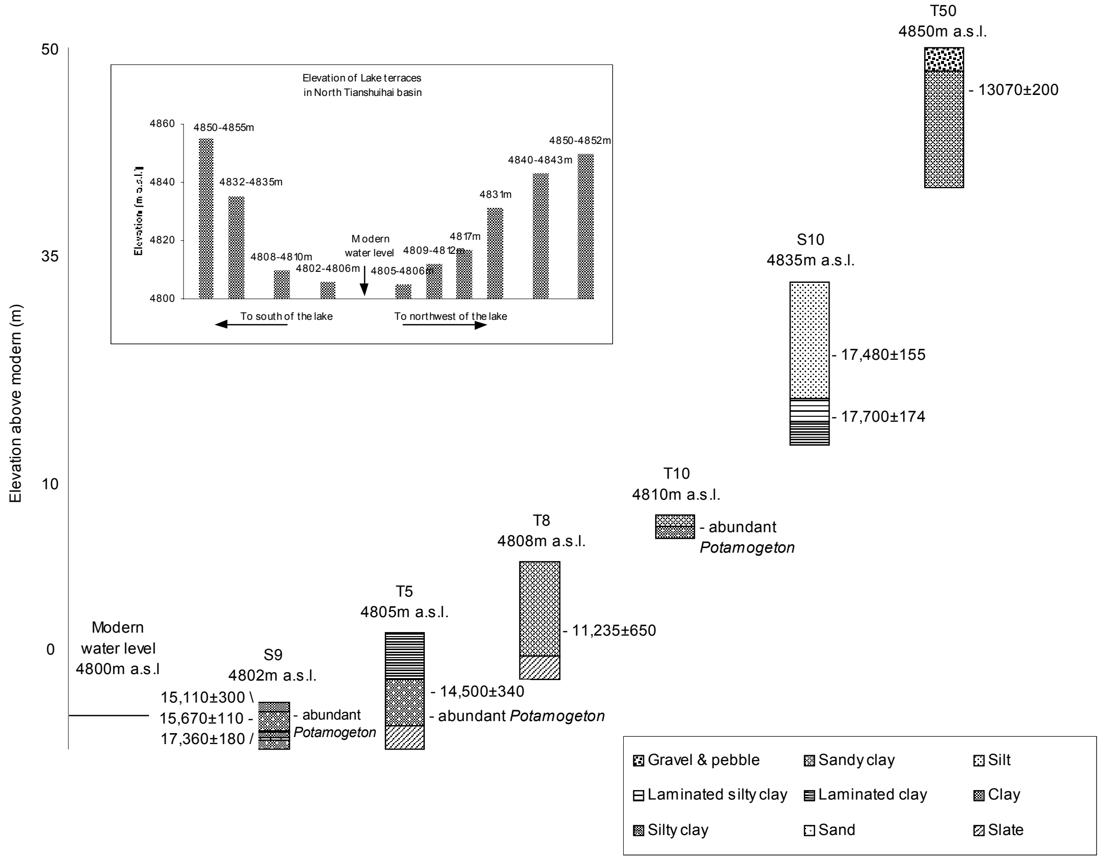


Coding

17,750-17,650 yr B.P.	very high (7)
17,650-17,450 yr B.P.	high (6)
17,450-17,400 yr B.P.	low (2)
17,400-16,345 yr B.P.	moderately low (3)
16,345-16,000 yr B.P.	low (2)
16,000-15,595 yr B.P.	moderately low (3)
15,595-15,520 yr B.P.	low (2)
15,520-15,370 yr B.P.	moderately low (3)
15,370-15,300 yr B.P.	intermediate (4)
15,300-13,750 yr B.P.	moderately low (3)
13,750-13,500 yr B.P.	high (6)
13,500-13,000 yr B.P.	extremely high (8)
13,000-12,800 yr B.P.	very high (7)
12,800-11,300 yr B.P.	moderately low (3)
11,300-11,000 yr B.P.	intermediate (4)
0 yr B.P.	very low (1)

Preliminary coding: 27-11-1998
Second coding: 8-1-1999
Final coding: 24-3-1999
Coded by GY and SPH

3.37. Zabuye Lake, Xizang (Tibet) Autonomous Region

Zabuye Lake (31.35°N, 84.07°E, 4421m above sea level) is a large bittern-salt lake in the Gangdisi Mountains in the interior of Tibetan Plateau. The lake consists of two sub-basins, a southern and a northern sub-basin, joined together by a relatively narrow channel. The lake has total area of 243 km^2, a mean depth of 0.7m and a maximum depth of less than 2m (Zheng et al., 1989). The lake has no outflow, but there are two inlets, one entering from the west and one from the east. The catchment area is 6680 km^2, and the basin is surrounded by mountains ca 4600-5200 m a.s.l. The lake is supported by surface runoff, underground water, and glacier meltwater which provides ca 2.87% of the inputs by volume (Zheng et al., 1989). The salt content in the lake water is 360-410g/L (Zheng et al., 1989). The basin originated through faulted structures. The underlying bedrock is Cretaceous-Eogene acidic igneous, mudstones and sandstones. The climate in the catchment is extremely arid, with 192.6mm total annual precipitation but 2341.6mm total annual evaporation, leading to a negative water balance throughout the year (Zheng et al., 1989). A large area of playa is being exposed around the lake, and mirabilite is currently being deposited in the lake. No aquatic plants grow in the lake, and only sparse *Stipa purpurea* occurs in the basin (Wu and Xiao, 1996).

The presence of lacustrine deposits at elevations above modern lake level indicate that the lake has been considerably higher than today in the past (Zheng et al., 1989; Qi and Zheng, 1995). These deposits have been studied at several sites within the basin, including at Chaduixiong, Jiadonglongba, Jiuer and Najiunihe. The original authors suggest that there are a number of lacustrine terraces and also lacustrine spits marking former shorelines. They group these features, on the basis of elevation, into 14 separate shoreline levels (L1-L14; see Table below).

Level	Terrace (m a.s.l.)	Spit (m a.s.l.)	Lake area (km^2)	Lake volume (10^8 m^3)	Mean depth of lake (m)	^{14}C (yr B.P.)
L1	4430	4432-4443	350	24.5	7	
L2	4440	4443-4455	480	68.6	14.3	5315±135 (southern profile) 9510±165 (northern profile)
L3	4446	4449-4461	550	101	18.3	
L4	n/a	4463-4467	720	199	27.6	
L5	4470	4471-4477	810	267	33	8725±135 (Chaduixiong)
L6	4492	4494-4496	1150	451	39.2	12,535±180 (Jiadonglongba), 22,670±380 (Jiuer)
L7	4500+	4508-4520	1350	575	42.7	23,770±600 (Chaduixiong)
L8	n/a	4532-4555	1381	665	48.2	
L9	4590-4600	4600	9780	11700	120	
L10	4621	n/a				
L11	4640-4642	n/a				
L12	4665	n/a				
L13	4686	n/a				
L14	4690-4706	n/a				

The terrace features are clearly formed from the erosion of lake-bottom sediments during intervals when the lake dropped. Some of the so-called spits also appear to contain lake-bottom sediments. This, and the fact that they appear to be linear features within former embayments of the lake, raise doubts about whether they are constructional shoreline features (as implied by their name). They could more plausibly be considered as underwater bars or berms. Furthermore, the elevation of the terraces and spits of a given age varies from site to site. Again, this makes it seem less likely that these features are tied to specific levels of the lake. We do not therefore believe that it is possible to use these lacustrine relicts to reconstruct lake levels explicitly. However, we use the presence of lacustrine material at elevations above modern lake level to infer a minimum water depth for specific radiocarbon-dated periods.

There are four terraces (L10-L14) at altitudes of >4600 m a.s.l. These terraces consist of lake-bottom sediments and appear to have been created by erosion. The presence of lacustrine deposits above 4600 m a.s.l. indicates that the lake was open and overflowing during the earliest part of its history (Zheng et al., 1989, 1996). Unfortunately, there are no detailed studies on these terraces and the lacustrine sediments are not dated.

The Zabuye Basin becomes closed at ca 4600 m a.s.l. Lacustrine deposits found below this elevation are therefore likely to have been formed in a closed lake. Calcareous sand and pebbles have been found in the so-called L9 spit at Chaduixiong at elevations of ca 4600 m a.s.l. These sediments appear to be lacustrine beach deposits. They are undated. Assuming that the lake level was ca 4600 m a.s.l. when these beach deposits were formed, the lake was ca 280m higher than today and the lake area was ca 9780km^2. A similar calcareous gravel and pebble unit has been found in the so-called L8 spit at Chaduixiong at an elevation of ca 4532m. The unit is undated. Assuming this unit corresponds to a beach deposit, the lake level was ca 111-121m higher than today and the lake area was ca 1831km^2. Sediments in the so-called L7 spit at Chaduixiong (elevation 4510 m a.s.l.) are yellow sandy clay with pebbles. The more clayey nature of these deposits suggest they are littoral zone deposits rather than beach deposits. These deposits are undated but indicate a lake level of >4510 m a.s.l.

Yellow sandy clay with pebbles, similar to the deposits found a Chaduixiong, were found at an elevation of 4510m at a site to the west of southern lake basin. The ostracode assemblage from these sediments is characterised by *Leucocythere postilirata,* characteristic of brackish-water lakes with a salt content of <10g/L (Zheng et al., 1989). Thus, the ostracode assemblage is consistent with the idea that these are nearshore sediments deposited in a lake that was considerably deeper (>90m) than present. A sample from the top of the profile was radiocarbon-dated to 23,770±600 yr B.P. suugesting that the deep-water lake was in existence before ca 23,800 yr B.P.

Light-yellow carbonate clay deposits were found in the so-called L6 spit at Jiuer. The carbonate-rich, clayey nature of this deposit suggests it was deposited in the nearshore zone of a lake. The top of this feature is 4492 m a.s.l. A sample from a depth of 4480m a.sl. was radiocarbon-dated to 22,670±380 yr B.P. Light-yellow carbonate clay deposits, characteristic of nearshore deposition, were found in the so-called L6 spit at Jiadongxiongba. The top of this feature is 4494 m a.s.l. A sample from a depth of 4485 m a.s.l. was radiocarbon-dated 12,535±180 yr B.P.

Zheng et al. (1989) assume that the features at Jiuer and Jiadonglongba, because of their approximate similarity in elevation (4490-4492m) and lithology, mark the shoreline position of a single lake. Using the two available radiocarbon dates from these profiles, they assume that this lake was present between 22,600 and 12,400 yr B.P. This is an unreasonable assumption, given that the so-called spits are clearly not shoreline features, and we therefore interpret these deposits as indicating lake levels > 4490m around 22,600 yr B.P. and >4492 m a.s.l. around ca 12,400 yr B.P. Given the fact that there is only a single radiocarbon date from each profile, it is not possible to estimate how long these lakes persisted.

Grey-white carbonate clays, typical of nearshore deposits, were found in the so-called L5 spit at Chaduixiong at an elevation of ca 4471 m a.s.l. The unit contains some grains of *Myriophyllum* pollen, which indicates relatively fresh-water conditions. The presence of *Cundimella mirabilis*, an ostracode species characteristic of brackish water with a salt content of 20-30 g/L (Zheng et al., 1989), is consistent with the other lines of evidence. A sample at 4470m from this profile was radiocarbon-dated to 8725±135 yr B.P., suggesting the lake level was >50m higher than present before 8700 yr B.P.

Carbonate clays are characteristic of the so-called L4 spit (elevation 4463m) at Chaduixiong. The presence of *Pediastrum* and abundant *Myriophyllum* pollen indicates that the water was fresh. The presence of *Leucocythere mirabilis*, an ostracode species characteristic of wide range of water conditions and salinity (Zheng et al., 1989), is consistent with the idea that the lake was reasonably fresh. The material is undated and the significance of this unit in terms of lake level is therefore unclear. Carbonate silty-clay and clay are also found in the so-called L3 spit (elevation 4449m) at Chaduixiong. The presence of more mineraogenic material may indicate that this unit was formed in slightly deeper-water conditions than the sediments in L4. Whether these deposits were formed at the same time as the deposits in L4 is unclear. Given that the material is undated, it is not possible to use these deposits to code lake-level changes.

Light-grey carbonate clay with sand occurs in a profile to the south of the northern sub-basin of the modern lake. The lithology suggests nearshore to littoral deposition. A sample from 4440m was radiocarbon-dated to 9510±165 yr B.P., indicating that lake level was >19m higher than today before ca 9500 yr B.P.

Light-grey carbonate clay with sand occurs in a profile to the west of the southern sub-basin of the modern lake. The lithology suggests nearshore to littoral deposition. A sample from 4440m was radiocarbon-dated to 5315±135 yr B.P., incicating that the lake levels was >19m higher than today around 5300 yr B.P.

Light-grey carbonate clays were found in the so-called L1 spit (elevation 4432m) at Chaduixiong. The ostracode assemblage was characterised by *Limnocythere dubiosa* and *Limnocytherellina binoda*, a species characteristic of saline water with salt content of 80g/L (Zheng et al., 1989). Zheng et al. (1989) interpret the L1 spit as indicating a phase when the lake was ca 11m higher than today sometime after ca 5300 B.P. It is equally plausible, given the difference in elevation and lithology, to assume that the deposits in L1 at Chaduixiong correspond to a deeper-water facies of the L2 deposits found there. Given the absence of radiocarbon dates from this profile, there is no way to determine the significance of these deposits with regard to lake levels.

A more continuous record of lacustrine deposition in the Zabuye basin is provide by cores from the lake floor. There are nine engineering cores from the lake, five (Cores CK1, 2, 3, 4, 5) of which were taken along a south-north transect across the southern sub-basin, three (Core CK6, 7 and 8) from the southern part of the northern sub-basin (Qi and Zheng, 1995; Zheng et al., 1996). The site of the remaining engineering core (Core F32) is unclear (Zheng et al, 1989). A 20.12m-long core (ZK91-2) was obtained from the playa to the west of the modern shoreline (Qi and Zheng, 1995). Only the records from CK1 and ZK91-2 have been published in detail. The cores provide sedimentary records back to ca 40,000 yr B.P. Changes in water depth are reconstructed from changes in lithology, geochemistry, aquatic pollen or algae and ostracode assemblages of these cores. The chronology is based on 4 AMS from Core ZK91-2 (Zheng et al., 1996; Wu and Xiao, 1996) and 13 conventional radiocarbon dates from Cores 1, 2, 4, 5, 8 and Core F32 (Zheng et al., 1989).

The basal sediment (20.12-19.79m, Core ZK91-2) is dark-yellow fine sand with small stones. The lithology suggests this is an alluvial deposit. The unit is undated, but extrapolation of the sedimentation rate (0.087 cm/yr) from between the basal two radiocarbon dates from the upper part of the core suggest it was formed before ca 37,600 yr B.P. Similar deposits are found overlying weathered bedrock in Core CK1 (below 10.5m), the material is undated but extrapolation of the sedimentation rate (0.0171 cm/yr) between the two radiocarbon dates from this core suggest that it was formed before 59,500 yr B.P. This estimate is likely to be unreasonable because the radiocarbon dates are rather far away; the extrapolated age from ZK91-2 seems more reasonable based as it is on a date of 29,330±420 yr B.P. from 12.54-12.84m.

The establishment of lacustrine conditions is registered by the deposition of carbonate clays in both cores (19.79-17.55m in Core ZK91-2; 10.50-9.14m in CK1). The unit contains fine silt in the marginal core (ZK91-2), which is consistent with the relative positions of the two cores. Extrapolation of the sedimentation rate calculated from the radiocarbon dates in the overlying units in ZK91-2 indicates that this unit was formed between 35,000-37,600 yr B.P., while the same procedure in core CK1 suggests it was formed between 51,500 and 59,500 yr B.P. Again, the estimate based on the CK1 core is unreasonable.

Increased water depth after ca 35,000 yr B.P. is indicated by the disappearance of silt from the carbonate clays in ZK91-2 (17.55-11.79 m) and reduced carbonate content (to <5%) in the deeper-water core CK1 (9.14-7.01m). The ostracode assemblage from CK1 is dominated by *Limnocytherellina kunlunensis, L. trispinosa* and *Leucocythere mirabilis*, species consistent with relatively fresh, deep-water conditions. A sample from near the upper boundary of this unit (12.54-12.64m) in ZK91-2 was AMS-radiocarbon-dated to 29,330±420 yr B.P., indicating that this moderately deep water lake occurred between 35,000-29,000 yr B.P. (Extrapolation based on the available radiocarbon dates in CK1 would date this unit to 39,000-51,500 yr B.P.).

Decreased water depth is indicated after ca 29,000 yr B.P. In the deepwater core, CK1 (7.01-4.21m), the carbonate content increases (to 10-20%). In the shallower-water core, ZK91-2 (11.79-7.5m) the carbonate clay contains gravel, and shows a number of structural features (e.g. ripples, water-scour marks) consistent with nearshore conditions. The presence of abundant *Pediastrum simplex* and Cyperaceae pollen is consistent with shallow conditions. The ostracode assemblage from CK1 is characterised by the presence of *Candona* spp. and *Limnocytherellina kunlunensis*; the appearance of *Candona* (a typical freshwater species) is not consistent with shallowing. By interpolation between the lowermost radiocarbon dates on the ZK91-2 core, this phase of decreased water depth occurred ca 23,500 to 29,000 yr B.P. A sample from the top of the unit in CK1 was radiocarbon-dated to 22,610±500 yr B.P. (at 4.2m). On the basis of this date, the phase of decreased water depth occurred before 22,700 yr B.P.

An increase in water depth is indicated by the development of banding in the overlying sediments in both cores (7.5-5.95m in ZK91-2; 3.26-4.21m in CK1). The carbonate content in the marginal core (ZK91-2) decreases (<5%), consistent with increased water depth. There is also a decrease in the abundance of *Pediastrum simplex* and an increase in *Pediastrum boryanum*, a planktonic species of green algae, again consistent with deepening. A sample from 6.25-6.35m in ZK91-2 was AMS-radiocarbon-dated to 22,130±235 yr B.P. Interpolation of the sedimentation rate (0.087 cm/yr) between this date and the underlying date, suggests the deep-water phase occurred between ca 23,500 and 20,900 yr B.P. A sample at 4.2m in Core CK1 was radiocarbon-dated to 22,610±500 yr B.P. On the basis of this date, the deep-water phase occurred ca 17,050 to 22,700 yr B.P.

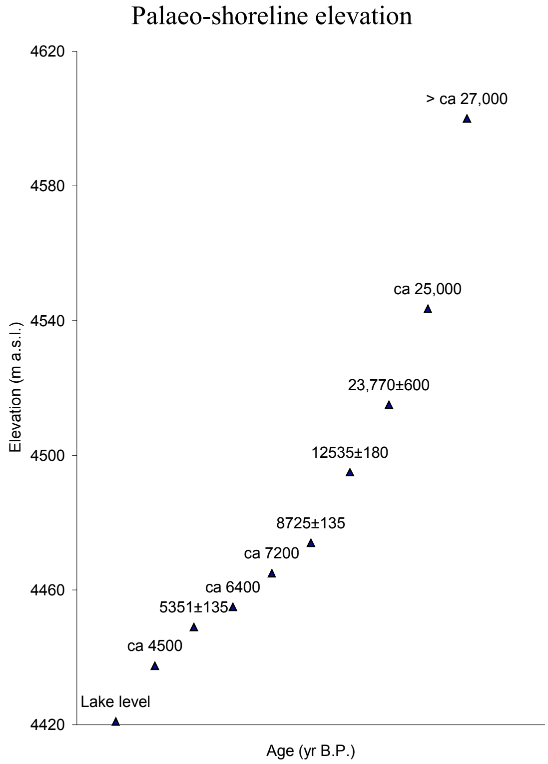
The disappearance of banding in the overlying sediments (5.95-3.70m in Core ZK91-2; 3.26-2.23m in CK1) in both cores indicate shallowing. The sediments in ZK91-2 are carbonate clay with fine silt. The presence of silt, the high calcium carbonate contents (40-50%) and the occurrence of dolomite are all consistent with shallowing. In CK1, the unit contains gravel and fine pebbles and the organic content is higher. Interpolation of the sedimentation rate (0.0275 cm/yr) between the AMS dates on over- and under-lying units in ZK91-2, suggests this shallow-water phase occurred ca 20,900-12,700 yr B.P. According to the dates on CK1, this shallow phase occurred between 11,000 and 17,050 yr B.P.

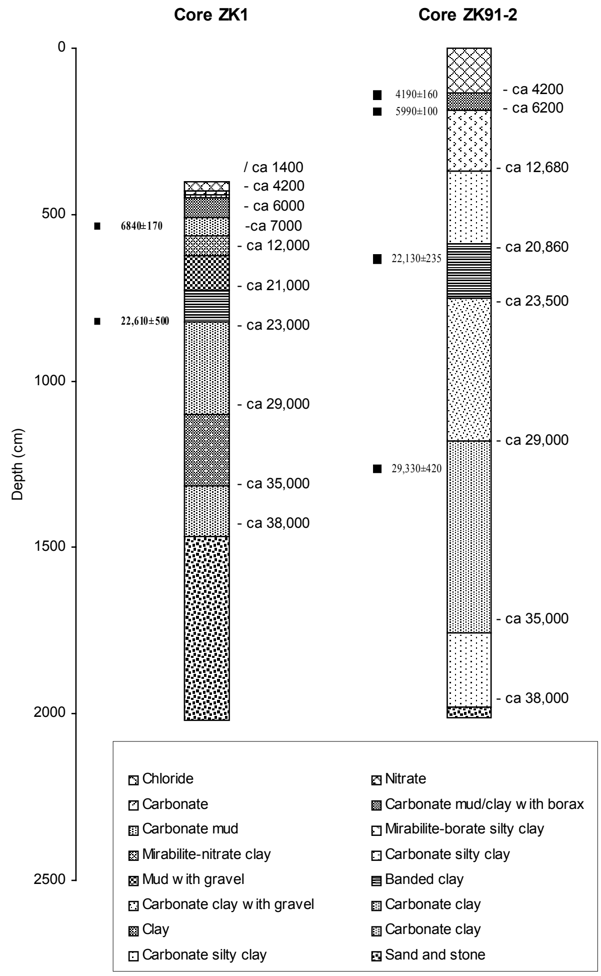
Further shallowing is indicated by chemical changes in the sediments. In the deepwater core (CK1, 2.23-1.63m) the nitrate content of the sediments increases markedly (20-30%). The presence of *Pediastrum* is consistent with shallow, saline conditions. Again, the ostracode record from the CK1 core, which is characterised by *Candoniella mirabilis* and *Candona spp.,* does not appear to be consistent with the other evidence. At the shallow-water coring site (ZK91-2, 3.70-1.87m) mirabilite-borax clays were deposited. The age of this unit is 7580-11,000 yr B.P. (according to the dates from CK1) and 6200-12,700 yr B.P. (according to the dates from ZK91-2).

A return to deeper-water conditions is indicated by the deposition of carbonate muds with low salt contents in both cores (1.87-1.36 in ZK91-2; 1.63-0.5m in CK1). The change in mineralogy suggests the lake water freshened and water depth increased. A sample from the lower boundary (1.81-1.97m) of the unit in ZK91-2 was AMS-radiocarbon-dated to 5990±100 yr B.P., suggesting this phase dates to ca 6200-4200 yr B.P. The estimated age according to the dates on CK1 would be 7580-2280 yr B.P. The evidence from CK1 and the other engineering cores suggest that there were fluctuations in water depth during this interval (not shown by the sediments in ZK91-2). Given the dating uncertainties of the CK1 core, and the impossibility (in the absence of information about depths of radiocarbon dates or unit boundaries) of correlating the other core records with the ZK91-2 record, we do not code these oscillations.

The uppermost sediment (1.36-0m in ZK91-2) is clay with high mirabilite and chloride salts. The change in mineralogy indicates the formation of a playa lake. A sample at 1.36m was AMS-radiocarbon-dated to 4190±160 yr B.P., indicating that the change to playa conditions occurred after ca 4200 yr B.P. In CK1 the deposits can be subdivided into three sub-units, showing a progressive in salt content from carbonate-rich clays (0.5-0.4m) through nitrate-rich clays (0.4-0.3m) to sediments with mirabilite and borax (0.3-0.0m).

In the status coding, very low (1) is alluvial deposits; low (2) by playa deposits associated with modern lake level; moderately low (3) saline clays in both cores; intermediate (4) by carbonate clays, with gravel and structural features typical of nearshore environments in marginal core; moderately high (5) by carbonate clay in the central core and carbonate silty clay in the marginal core; high (6) by low-carbonate clays (with no silt) in both cores; very high (7) by banded clays in both cores.


Palaeo-shoreline elevation
4620
4580
4540
4500
4460
4420
Elevation (m a.s.l.)
Age (yr B.P.)
> ca 27,000
ca 25,000
23,770±600
12535±180
8725±135
ca 7200
ca 6400
5351±135
ca 4500
Lake level


Core ZK1
Core ZK91-2
0
500
1000
1500
2000
2500
Depth (cm)
6840±170
22,610±500
/ ca 1400
- ca 4200
- ca 6000
-ca 7000
- ca 12,000
- ca 21,000
- ca 23,000
- ca 29,000
- ca 35,000
- ca 38,000
4190±160
5990±100
22,130±235
29,330±420
- ca 4200
- ca 6200
- ca 12,680
- ca 20,860
- ca 23,500
- ca 29,000
- ca 35,000
- ca 38,000
Chloride
Nitrate
Carbonate
Carbonate mud/clay with borax
Carbonate mud
Mirabilite-borate silty clay
Mirabilite-nitrate clay
Carbonate silty clay
Mud with gravel
Banded clay
Carbonate clay with gravel
Carbonate clay
Clay
Carbonate clay
Carbonate silty clay
Sand and stone

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Radiocarbon dates

29,330±420	12.54-12.64m, calcareous clay, Core ZK91-2, AMS date
23,770±600	4510 m a.s.l., calcareous clay, Chaduixiong
22,670±380	4480 m a.s.l., carbonate, Jiuer
22,610±500	ca 4.2m, carbonate, Core CK1
22,130±235	6.25-6.35m, clay, Core ZK91-2, AMS date
20,080±450	ca 5.5m, carbonate, Core CK2
18,620±300	ca 3.1m, carbonate, Core CK2
*12,535±180	4485 m a.s.l., calcareous clay, Jiadonglongba
* 9510±165	4440 m a.s.l., carbonate, northern profile
*8725±135	4470 m a.s.l., carbonate, Chaduixiong
6840±170	ca 1.5m, carbonate, Core CK1
5990±100	1.81-1.91m, clay, Core ZK91-2, AMS date
5980±80	ca 4.5m, mud, Core CK4
5770±80	ca 2.1m, mud, Core CK4
*5315±135	4440 m a.s.l., carbonate, southern profile
4470±80	ca 1.6m, calcareous clay CK8
4190±160	ca 1.36m, mirabilite clay, Core ZK91-2, AMS date
3950±80	ca 2.0m, mud, CK5
3530±70	ca 0.4m, calcareous clay, CK8, probably ATO
3150±70	ca 1.20m, calcareous clay, CK8
3150±70	ca 0.8m, carbonate, CK4
2170±150	ca 0.2m, borax-carbonate, F32
1350±70	ca 0.25m, chlorate-mirabilite, Core CK4

(4 samples were AMS-dated by the NSF-AMS Lab of Alexandria University; * means dates from Institute of Seismology and Geology of China; the rest of the dates are from Institute of Vertebrate Palaeontology, Chinese Academy of Science.)

Coding

--- 37,600 yr B.P.	very low (1)
37,600-35,000 yr B.P.	moderately high (5)
35,000-29,000 yr B.P.	high (6)
29,000-22,700 yr B.P.	intermediate (4)
23,500-17,050 yr B.P.	very high (7)
20,900-11,000 yr B.P.	moderately high (5)
12,700-6200 yr B.P.	moderately low (3)
6200-4200 yr B.P.	high (6)
4200-0 yr B.P.	low (2)

Preliminary coding: 08-11-98
Second coding: 30-11-1998
Third coding: 17-12-1998
Fourth coding: 22-03-1999
Final coding: 27-01-2001
Coded by GY and SPH

3.38. Zhacang Caka, Xizang (Tibet) Autonomous Region

Zhacang Caka (32.60°N, 82.38°E, 4328m above sea level) is a large playa basin in the interior part of the Tibetan Plateau (Zheng et al., 1989). Within the large playa area of Zhacang Caka, there are three salt lakes (Lakes I, II and III). These lakes have areas of 23.25 (Lake I), 57.5 (Lake II) and 29.0km^2 (Lake III) respectively and depths between 0.3-2m. Lakes I and II are separated by a +4-5m-high and 200-300m wide terrace. Lakes II and III are separated by a +20-30m-high and 300-500m wide terrace. The origin of these terraces is unclear although, as they consist of lacustrine carbonate deposits, we assume that they are relicts of a formerly extensive lake in the basin and that the deposits were incised when the modern lakes were formed. A number of recent beach ridges, associated with regression phases of the modern lakes, overlie the older terrace deposits. The three lakes are hypersaline; the salt content of Lake II, for example, is 296.91 g/L (Zheng et al., 1989). The modern lake deposits are mirabilite-mud or marl in water depths of 1-2m, muddy or clayey halite in water depths of 0.5-1m, and halite in water depths < 0.5m (Zheng et al., 1989). Two rivers (Pawacangbu and Xingqiannongbu) enter the Zhacang Caka from the southern high mountains (5000-6000 m a.s.l.), and there are some springs associated with Lake I and II (Zheng et al., 1989). The Zhacang Caka basin originated through faulting, but there is no evidence of recent tectonism. The underlying bedrock is Mesozoic sandstone and granite. The climate in the catchment is cold (-0.2°C annual mean temperature) and arid with 151mm total annual precipitation but 2303mm total annual evaporation (Zheng et al., 1989).

Twenty-three cores (maximum depth up to 15m) have been taken from the three salt lakes in order to investigate salt mineral resources (Zheng et al., 1989). The stratigraphy of the cores is similar (Zheng et al., 1989). The basal sediments are lacustrine yellow-brown carbonate clay, containing 1 or 2 layers of sandy silt or gravel sand. The thickness of this unit is >3.6m. Zheng et al. (1989) interpret the carbonate clay as indicating a fresh, deepwater lake. The presence of sandy silt or gravel sand units within the carbonate clays indicates minor oscillations towards shallower (though still fresh) conditions. Zheng et al. (1989) estimated this freshwater phase occurred before ca 15,000 yr B.P. The second sediment unit is black-blue marl or black muddy clay with a salt content of 5-10%. The salt minerals include gypsum, mirabilite and some halite. The thickness of this unit is between 4.8-8.6m. The more organic nature of the sediments (indicated by the colour) and the presence of evaporite minerals indicates that these deposits represent a shallower water phase. Zheng et al. (1989) estimated this more saline, shallow-water phase occurred in the late glacial and early to mid-Holocene. The uppermost unit consists of pale-white or gray salt deposits. The unit contains between 4-35% carbonate, 50-90% mirabilite, and 4-35% halite. The unit has a thickness between 1-5m and was deposited during the late Holocene (Zheng et al., 1989).

Although there are 20 published radiocarbon dates on 10 cores (76CK1, 76CK2, 76CK3, 76CK4, 76CK5, 76CK7, 76CK9, 76CK11, 78CK2, and 78CK3), detailed stratigraphic records have been published for only two of the cores, namely 78CK3 (Huang et al., 1980) and 78CK2 (Zheng et al., 1989). The 15.8m-long core (Core 78CK3) taken from Lake III (Huang et al., 1980) provides a detailed sedimentary record back to ca 20,000 yr B.P. The 15.0m-long core (Core 78CK2) taken from Lake II (Zheng et al., 1989) provides a record back to ca 26,000 yr B.P. The two cores show a broadly similar stratigraphy. There are 5 radiocarbon dates from 78CK3 but only 2 radiocarbon dates from 78CK2 (Huang et al., 1980), so the chronology of changes in 78CK2 is less tightly constrained than that for 78CK3. Changes in relative lake depth are reconstructed on the basis of a synthesis of the changes in lithology and geochemistry in both cores. The differences in the dating control between the two cores necessarily introduces some uncertainties into the reconstruction of the timing of the changes. Neverthless, we are able to reconstruct a reasonably clear picture of lake evolution during the Late Quaternary.

The basal sediment in Core 78CK2 (15.0-13.6m) is pale-yellow lacustrine clay. The lithology indicates deep water conditions. The ratio of Ca/Mg is 5 (the maximum value recorded in the core). Such a high ratio implies freshwater conditions. The unit was not radiocarbon dated, although extrapolation of the sedimentation rate between the two radiocarbon dates on the overlying sediments suggests it was formed before 23,000 yr B.P. The unit is not found in Core 78CK3, which only covers the period from ca 20,000 yr B.P.

The overlying unit (15.8-14.27m in Core 78CK3, 13.6-12.2m in Core 78CK2) is a lacustrine clay containing sand and gravel. The increase in coarse material indicates nearshore deposition and suggests that the lake became shallower. The ratio of Ca/Mg is 3-4 (measured in Core 78CK2), suggesting that the lake remained relatively fresh. Nevertheless, the decrease in this ratio is consistent with shallowing. A sample from this unit at the base of Core 78CK3 (15.6-15.8m) was radiocarbon-dated to 20,000±350 yr B.P. However, Core 78CK3 does not necessarily penetrate to the base of this unit. A sample from the base of the overlying unit was radiocarbon-dated to 15,400±160 yr B.P. Thus, on the basis of information from Core 78CK3, this unit was deposited between 20,000-15,400 yr B.P. The evidence from Core 78CK2 suggests the unit is older. Thus, extrapolation of the sedimentation rate between the radiocarbon dates on overlying units suggests that the clay with sand and gravel unit was formed between 23,000 and 20,000 yr B.P.

The overlying unit is a carbonate clay and only appears in Core 78CK2 (12.2-12.0m). The finer nature of the sediments suggests an increase in water depth. The ratio of Ca/Mg is 3.0, similar to that of the underlying unit. The absence of this unit in Core 78CK3 suggests that there may have been loss of sediment through erosion, and may imply that the date of 15,400 yr B.P. is a maximum age for the end of the deposition of the clay with sand and gravel unit.

The overlying unit (14.27-7.87m in 78CK3, 12.0-4.8m in 78CK2) is variously described as a black muddy clay (78CK3) and a black marl (78CK2). The unit contains no evaporites. The change in colour probably indicates increased organic content, and suggests that the lake became slightly shallower. The ratio of Ca/Mg is between 3.0-4.0, consistent with relatively freshwater conditions. Two samples from 9.6-9.9m and 6.3-6.5m in Core 78CK2 were radiocarbon-dated to 15,600±600 and 9060±120 yr B.P., suggesting this unit was formed between 19,990 and 6790 yr B.P. However, two samples from 14.2-14.4m and 13.2-13.4m in Core 78CK3 were radiocarbon-dated to 15,400±160 and 13,400±160yr B.P. respectively, suggesting the unit was formed between 15,400-8020 yr B.P. We note that the age of 15,400 yr B.P. is a minimum age for deposition of this unit, given that there is likely a sedimentary hiatus between this and the underlying unit in Core 78CK3. The ages for the uppermost boundary of the unit (8020 yr B.P. and 6790 yr B.P.) are in relatively good agreement, given the limited numbers of radiocarbon dates available from both cores.

The next three units are only registered in Core 78CK3. The first (7.87-7.61m) is clayey mirabilite. Mirabilite comprises 90% of the total sediment. The marked increase in salt content indicates saline water conditions and a pronounced shallowing. By interpolation of the sedimentation rate (0.1009 cm/yr) between a radiocarbon date of 13,400±160yr B.P. on the underlying clay unit and a radiocarbon date of 4780±180 yr B.P. on one of the overlying units, this mirabilite-rich unit is dated to between 8020-7760 yr B.P.

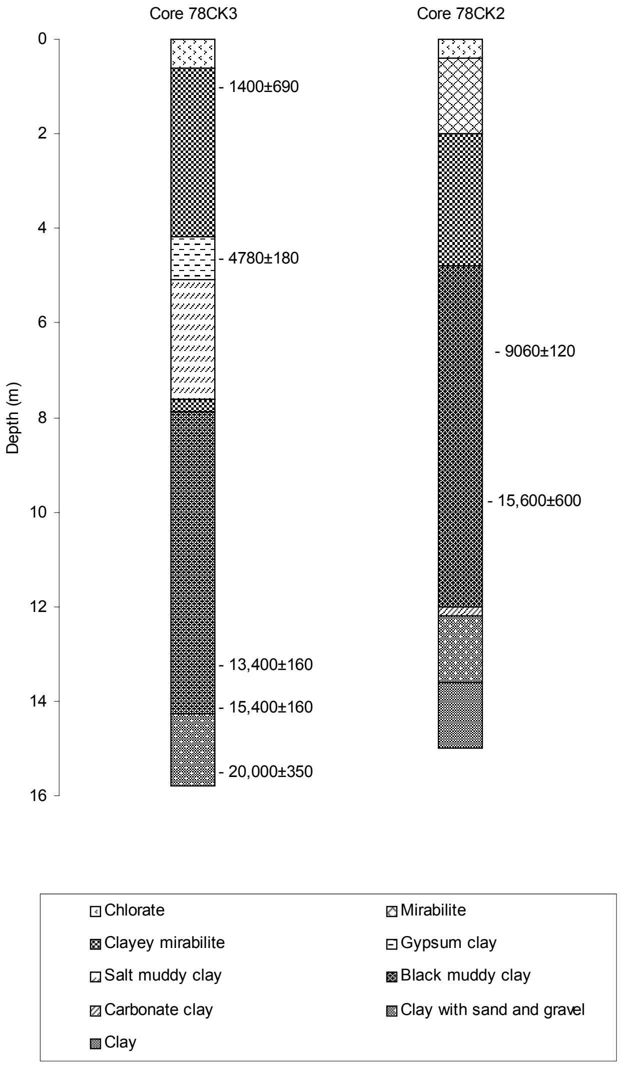
The overlying sediment (7.61-5.09m) is salt-bearing muddy clay, with 10% halite. The increase in the proportion of detrital minerals (90%) suggests decreased salinity and increased water depth. By interpolation, this unit is dated to between 7760-5270 yr B.P.

The overlying sediment (5.09-4.19m) is gypsum- and halite-rich clay. The abundance of evaporite minerals increased to 40% (10% gypsum and 30% halite), indicating increased salinity and decreased water depth. A sample from 4.5-4.7m was radiocarbon-dated 4780±180 yr B.P., suggesting the unit was formed between 5270-4340 yr B.P.

The overlying sediment, which is found in both cores (4.19-0.6m in Core 78CK3, 4.8-0.4m in Core 78CK2) is clayey mirabilite. The mirabilite content in 78CK3 varies from 100% near the base of the unit to ca 85% higher up. This is consistent with the fact that minor amounts of gypsum and halite are noted as being present within the unit in Core 78CK2. The increased salt content indicates increased water salinity and a significant decrease in water depth. The ratio of Ca/Mg (measured in Core 78CK2) is <1.0, consistent with saline conditions. A sample from 1.35-1.55m in Core 78CK3 was radiocarbon-dated to 1400±690 yr B.P. Interpolation of the sedimentation rate (0.093 cm/yr) between this date and the date on the underlying unit, suggests this saline phase occurred between 4340-580 yr B.P.

The uppermost sediment (0.6-0m in Core 78CK3, 0.4-0m in Core 78CK2) is variously described as clayey halite (78CK3) and mirabilite-bearing halite (78CK2). The presence of halite suggests extremely saline and very shallow conditions after ca 580 yr B.P. The ratio of Ca/Mg is <1.0, consistent with saline conditions.

In the status coding, extremely low (1) is indicated by the modern halite salt crust; very low (2) by clayey mirabilite; low (3) by gypsum- or halite-bearing clays; intermediate (4) by salt-bearing muddy clay with 10% halite; moderately high (5) by lacustrine clay with sand and gravel but containing no evaporites; high (6) by black clay or marl with no evaporites; very high (7) by carbonate clay with no evaporites; and extremely high (8) by lacustrine clays with no evaporites.



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Radiocarbon dates

20,000±350	15.6-15.8m, clay with sand and gravel, Core 78CK3
15,600±600	9.6-9.9m, muddy clay, Core 78CK2
15,400±160	14.2-14.4m, muddy clay, Core 78CK3
13,400±160	13.2-13.4m, muddy clay, Core 78CK3
10,900±200	4.65-4.85m, black mud, 76CK1
10,400±250	5.15-5.35m, black mud, 76CK9
9060±120	6.3-6.5m, muddy clay, Core 78CK2
8090±130	5.45-5.70m, black mud, 76CK11
8000±130	2.40-2.70m, black mud, 76CK1
7800±210	3.80-4.10m, mirabilite-bearing black mud, 76CK3
7000±110	4.80-5.00m, black mud, 76CK4
6500±160	3.65-3.88m, black mud, 76CK2
6200±160	3.90-4.10m, black mud, 76CK5
6070±210	2.30-2.55m, black mud, 76CK3
5710±130	2.78-2.99m, mirabilite-bearing mud, 76CK7
5600±150	1.74-1.94m, black mud, 76CK1
4780±180	4.5-4.7m, gypsum- and halite-rich clay, Core 78CK3
3840±130	2.47-2.70m, gray mud, 76CK11
3000±810	1.50-1.65m, black mud, 76CK2
1400±690	1.35-1.55m, salt-rich clay, Core 78CK3

(The samples were dated by the ^{14}C Lab of Seismology and Geology Institute, the State Bureau of Seismology).

Coding

26,000-23,000 yr B.P.	extremely high (8)
23,000-20,400 yr B.P.	moderately high (5)
20,400-20,000 yr B.P.	very high (7)
20,000-7000 yr B.P.	high (6)
8020-7760 yr B.P.	very low (2)
7760-5270 yr B.P.	intermediate (4)
5270-4340 yr B.P.	low (3)
4340-580 yr B.P.	very low (2)
580-0 yr B.P.	extremely low (1)

Preliminary coding: 4-2-1999
Second coding: 12-10-2000
Final coding: 18-11-2000
Coded by GY and SPH

3.39. Zigetangcuo, Xizang (Tibet) Autonomous Region

Zigetangcuo (Zigetang Lake, 32.08°N, 90.83°E, 4560m above sea level) is a large closed lake in the Ando Region of the inner part of the Tibetan Plateau (Shen and Xu, 1994). The modern lake has an area of 184 km^2. There are low-lying lacustrine plains, ca 3-4 km wide, to the south and north of the lake. The catchment area is ca 10-15 times the lake area (Shen and Xu, 1994). The basin is surrounded by mountains which reach elevations of 4900-5300 m a.s.l. The lake is fed mainly by direct precipitation and runoff from the basin. Three rivers (Simaijiuqu from the east, Zirongzangbu from the north and Bentoerqu from the south) drain from the high mountains around the basin into the lake. These rivers are fed by precipitation, springs and snowmelt (Shen and Xu, 1994). There is no outflow from the lake and the lake water is very saline (266.5 g/L). The Zigetang basin originated through faulting during the early Tertiary. Although tectonic activity and uplift have continued during the Quaternary, the impact of tectonism on the Zigetang basin during the last ca 35,000 years is negligible. The bedrock in the basin is Eogene red sandstone and mudstone. Subaerial sediments derived from weathering of the Jurassic bedrock have a characteristic red colouration and this colour is also characteristic of modern shallow water lacustrine deposits (Shen and Xu, 1994). The climate in the Ando region is cold (-3°C mean annual temperature) and dry (411.6mm total annual precipitation, 1770mm annual evaporation) (Shen and Xu, 1994). The Ando Region is characterised by alpine steppe and tundra vegetation, dominated by *Stipa* spp.

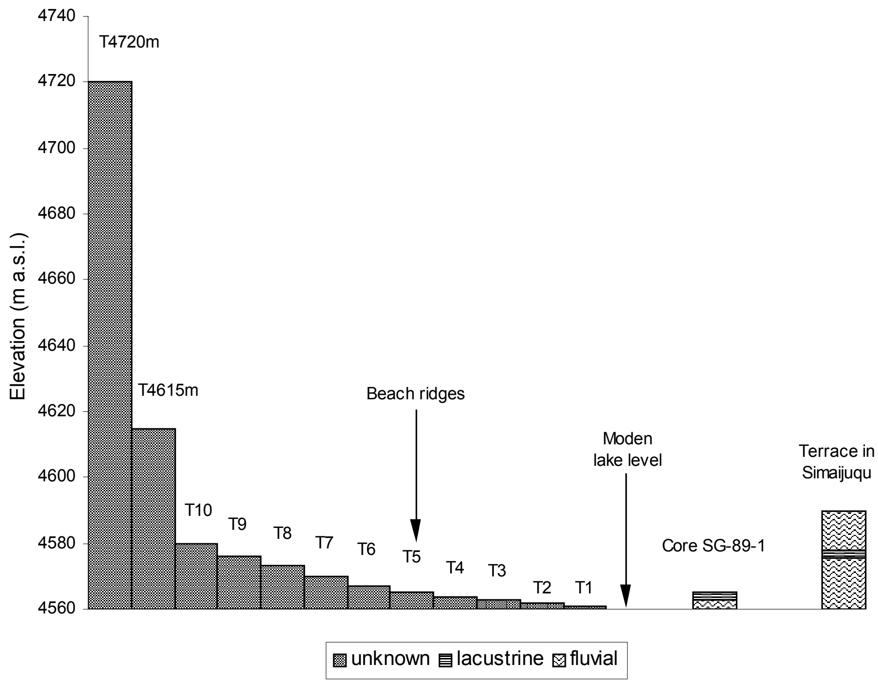
There are two lacustrine terraces to the south of the modern lake (Shen and Xu, 1994). The higher terrace, which occurs as discontinuous fragments, is at 160m above modern lake level (i.e. at 4720 m a.s.l.). The top of the lower terrace occurs at 15m above modern lake level (i.e. at 4615 m a.s.l.). The deposits comprising these terraces have not been described or radiocarbon dated. The top of the higher terrace occurs at a similar elevation to the highest terrace in the Cuona basin, 100 km east of Zigetangcuo. On the basis of this apparent congruity of elevation, Shen and Xu (1994) claim that the two lake basins were joined to form a single mega-lake sometime before 35,000 yr B.P. However, given the distance between the two basins, the absence of evidence of a significant spillway through the divide between the two basins, and the fact that there is no direct evidence to show that the two sets of terraces were joined, it seems unlikely that Zigetangcuo and Cuano were part of a single lacustrine system in the Late Quaternary.

A sequence of ten beach ridges occur to the south side of the modern lake. These ridges extend from modern lake level to an elevation of 4580 m a.s.l. The individual beach ridges occur at ca 4561m, 4562m, 4563m, 4564m, 4565m, 4568m, 4570m, 4573, 4576m and 4580m respectively. The highest of the ridges is ca 1km away from the modern lake margin. The structure of one of these beach ridges (T5, 4565 m a.s.l., ca 450m from the modern lake margin) was investigated by coring. The core (SG-89-1) is 2.6m long and shows a sequence from disturbed muddy clays (2.1-2.6m), through coarse gravels (2.0-2.1m) and into lacustrine sands and coarse gravels (0-2.0m). This sequence likely represents a transition from nearshore deposits into beach deposits. Although these ten beach ridges provide a geomorphic record of formerly higher lake levels (Shen and Xu, 1994), none of them are radiometrically dated. Shen and Xu (1994) claimed that the beach ridges were formed during the last 12,000 years as the lake became progressively lower. The idea that the beach ridges formed during the last 12,000 yr is based on the fact that they are younger than the lower 4615m terrace and the assumption that this 4615m terrace is the same age as the T2 terrace in the Cuona basin. Shen and Xu (1994) estimate that the top of the T2 terrace in the Cuona basin dates to ca 12,000 yr B.P. on the basis of extrapolation of the sedimentation rate between two radiocarbon samples both of which have ages of >20,000 yr. Given the fact that the upermost sediments in the T2 terrace are sandy beach deposits, this extrapolation seems unwarranted. Furthermore, it seems unlikely that the lower terrace in Zigetangcuo can be directly correlated with the T2 terrace in the Cuona basin. Thus, the ten beach ridges cannot be used to reconstruct the temporal sequence of lake-level changes in the Zigetangcuo basin.

The Simaijuiqu river enters the east side of the modern Zigetang Lake. The modern river has created a natural fluvial terrace through downcutting, presumably in response to a lowering of the lake (and hence of fluvial base level). The top of this terrace is at ca 4590 m a.s.l. (i.e. ca 30m above modern lake level and 18m above the river at that point). The 18m-deep section consists of fluvial sands below 14.5m. However, the overlying unit (12.0-14.5m, ca 4575-4578 m a.s.l.) consists of thinly-bedded lacustrine silty clay. The uppermost sediments (0-12.0m) consist of coarse sand and gravels of fluvial origin (Shen and Xu, 1993; Shen and Xu, 1994). The presence of a lacustrine unit within these predominantly fluvial deposits indicates that the lake was at least 15-18m higher than today. Samples the lower (14.5m) and upper (12.0m) boundaries of the lacustrine unit were radiocarbon-dated to 19,220±387 yr B.P. and 17,707±405 yr B.P. respectively, suggesting this high lake level occurred between ca 19,200-17,700 yr B.P.

The modern lake level (4560m a.s.l) is thought to be the lowest in the history of the lake (Shen and Xu, 1994).

In the status coding, low (1) is indicated by the modern lake level at 4560m; and high (3) by thinly-bedded lacustrine silty clays ca 15-18m above the modern lake. The beach ridge deposits and the lacustrine terraces are not coded because they cannot be dated with any reliability.



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Shen YP, Xu DM (1994) Fluctuations of lakes and their environments since last glaciation in Amdo Area, Tibet. Journal of Glaciology and Geocryology 16(2): 173-180 (in Chinese)

Radiocarbon dates

19,220±387	14.5m, silty clay, River Simaijiuqu profile
17,707±405	12.0m, silty clay, River Simaijiuqu profile

(The samples were dated in Lanzhou Institute of Glaciology and Geocryology, Chinese Academy of Science)

Coding

ca 19,200-17,700 yr B.P.	high (3)
0 yr B.P.	low (1)

Preliminary coding: 19-11-1998
Second coding: 21-12-1998
Final coding: 26-07-2000
Coded by GY and SPH

3.40. Erhai, Yunnan Province

Erhai (25°25'-26°16'N, 99°32'-100°27'E, 1974 m a.s.l.) lies in the western part of the Yunnan Plateau. The lake is elongated north-south, with a length of ca 42.58 km and a mean width of 5.85 km and an area of 249 km². The lake water is fresh. The maximum depth is 20.7 m and the mean depth is 10.2 m. The southern part of the lake is comparatively shallow. The lake is fed by the Miju River from the north and the Boluo River the south of the basin respectively, and also by more than ten smaller streams. The only output is via the Xier River into Rancangjiang River. The catchment has an area of ca 2565 km². The basin is of tectonic origin and was formed in the Early Pleistocene. The underlying bedrock in the western part of the basin is pre-Cambrian metamorphics, and Ordovician sandstone, shale and carbonates outcrop in the eastern part of the basin (NIGL et al., 1989; Zhang et al., 1998; Zhang, 1999).

There are no continuous lacustrine terraces around Erhai. However, isolated lacustrine deposits indicating lake levels above the elevation of the modern lake are found in several places. Isolated snail deposits typical of nearshore environments occur ca 1.5 km away from the west coast of the modern lake and at an elevationof ca 2-3m above modern lake level (NIGL et al., 1989). These snail shells were radiocarbon dated to 470±150 yr B.P. (GC-631), suggesting a phase of increased (+2-3m) water depth at ca 470 yr B.P. A sand bar (Shacun sand bar) occurs along the west coast of the modern lake. It lies ca 1-2m above modern lake level.Three snail samples from this sand bar were radiocarbon dated to 2230±90 (GC-1177), 1840±100 (GC-634) and 1250±80 (GC-1072) yr B.P. respectively, suggesting the lake was slightly higher (+1-2m) than its modern level during the period of 2230-1250 yr B.P. (NIGL et al., 1989). Finally, a peat deposit from the foot of the mountains in the western part of the basin (over +36 m above modern lake level, ca 4 km away from the west coast of the modern lake) has been radiocarbon dated to 3650±150 yr B.P. (GC-630) (NIGL et al., 1989). Although this peat deposit does not relate directly to changes in lake level, it does suggest a phase of more humid conditions occurred around 3650 yr B.P.

There are several cores providing information about the longer-term evolution of the Erhai basin (NIGL et al., 1989). Four long cores (core ZK14, ZK18, ZK26 and ZK27), taken from the southern part of the basin, provide a record covering the last glaciation (NIGL et al., 1989). Core ZK14 lies southwest of the southern end of the lake, near Xiaguang, at an approximate elevation of 2000 m a.s.l. (+26m above modern lake level). Cores ZK18 and ZK27 lies directly to the south of the lake, at an approximate elevation of 2010 m a.s.l. (+36m above modern lake level). Core ZK26 lies ca 10km east of the southern part of the lake in Fengyi County at an approximate elevation of 2020 m a.s.l. (+46m above modern lake level). There are 7 radiocarbon dates on these cores (NIGL et al., 1989). Since the cores occur between 2-10 km beyond the margin of the modern lake and at higher elevations, the presence of lacustrine deposits in the cores indicates the existence of a considerably larger lake than today.

The sediments of core ZK 26 (24m long) consist of a sequence from basal sand (22-24m, lacustrine clay (21.5-22m), sand (20-21.5m), lacustrine clay (15-20m), gravel (14-15m), sand (10.5-14m), lacustrine clay (9.5-10.5m), gravel (6-9.5m) and lacustrine clay (0-6m). This sequence shows four phases of much higher-than-present lake levels (corresponding the the lacustrine clay deposits) with intervals of fluvial/colluvial activity between (indicated by the gravel and sand units). Unfortunately, only one of the lacustrine intervals has been dated. A single sample of charcoal from a depth of 20.29-21.02 m was radiocarbon-dated to 34,090±3800 yr B.P. This date indicates an early phase of expanded lake conditions.

The lacustrine sediments in core ZK14 (18m long) consist of sequence from silty clay (17-18m), clay (16-17m), sandy clay (10-16m), silty clay (3-10m), sand (2-3m) and silty clay (0-2m). Two samples from 16.08-16.28 m and 4.64-4.94 m are radiocarbon dated to 23,050±1300 and 21,650±830 yr B.P. This suggests that Erhai Lake was much larger than today the deepwater phase recorded as occurring from sometime before ca 23,000 to sometime after 21,600 yr B.P.

The basal unit (ca 44-36 m) of core ZK 18 (44m long) is gravel. A sample from a thin layer of organic clay interbedded in this unit is radiocarbon dated to 22,265±1070 yr B.P. The overlying unit (ca 36-29 m) is sand. These basal units are non-lacustrine. The presence of non-lacustrine deposits at this higher-elevation site, during the interval marked by lacustrine deposition in core ZK14, places a constraint on the maximum extension of Erhai Lake during this interval. The overlying unit (24-29m) is lacustrine silty clay. The presence of lacustrine material in this higher-elevation core indicates a significant increase in lake level. A sample from the bottom of this unit is radiocarbon dated to 18,700±560 yr B.P., suggesting this increase in lake level began ca 18,700 yr B.P. The overlying units, which are not dated, consist of a sequence of lacustrine clay (20-24m), silt clay (15-20m), sand (12-15m), clay (10.5-12m) and silty clay (0-10.5m). These deposits indicate continued lacustrine or peri-lacustrine deposition at the site.

The sequence of deposits from core ZK 18 is repeated in core ZK27. Two radiocarbon dates on the uppermost part of this sequence in core ZK27 enable us to provide a more precise chronology for the younger part of the record. Thus, the sediments of core ZK 27 (16m long) consist of a sequence from basal gravels, to lacustrine silty clay, clay (11.2-12m), sandy clay (8-11.5m), clay (4-8m) and silty clay (0-4m). A sample (11.7-11.3 m) from the lowermost clay unit has been radiocarbon-dated to 17,030±510 yr B.P. A second sample from the uppermost silty clay unit (3.4-2.9m) has been radiocarbon dated to 11,610±300 yr B.P. These dates imply that the expanded lake phase indicated by the sediments in cores ZK 18 and 26, persisted during the interval from 18,7000 to sometime after 11,600 yr B.P.

A record of Holocene sedimentation in the basin is provided by 6 cores (core Er1, Er2, core 4, core 48 and two unnamed cores) from the lake-bottom sediments (NIGL et al., 1989). Four of these cores (core Er1, core Er2, core Er4 and core Er48) have been studied in more detail. Core Er1 (2.8 m long) and core Er2 (3.1 m long) were taken very close together, in a water depth of 4.6m offshore from Jinsuo Island in the southern part of the lake. They show similar lithology and have been described as a single record (Er). This combined record provides lithological, mollusc and diatom data back to ca 11,300 yr B.P. (Zhang et al., 1998; Zhang, 1999). Core Er4 (1 m long), was taken from a water depth of 4.3m in the northern part of the modern lake. Core Er48 (1.6 m long) was taken from a water depth of 8.7m in the southern part of the modern lake. The lithological descriptions of these two cores are less detailed than those of the main Er core, and there is no quantitative information about changes in the aquatic assemblages. The two unnamed cores, for which we have no lithological information but which provide two additional radiocarbon dates, were taken in shallow water (<2m deep) from the western bay of the lake near Shacun sand bar (NIGL et al., 1989). There are a total of 8 radiocarbon dates on the cores from the lake-bottom deposits (NIGL et al., 1989; Zhang et al., 1998; Zhang, 1999). Changes in water-depth are reconstructed from the changes in lithology and diatom assemblage in the cores, and broadly follow the interpretation of the original authors (Zhang et al., 1998; Zhang, 1999).

The basal unit (310-288 cm) in core Er is dark grey, clayey silt and silt. The lithology is consistent with moderately deep lacustrine conditions. The diatom assemblage is dominated by planktonic *Cyclotella* sp. (40-50%) (Zhang, 1999), again consistent with moderately deep water. There is no radiocarbon date from this unit. However, extrapolation of the sedimentation rate (0.28 mm/yr) between the two radiocarbon dates on overlying units (7754±45 yr B.P from 1.98-2.03 m and 5825±40 yr B.P. from 1.44-1.48 cm) suggests this deepwater phase occurred ca 11,680-10,900 yr B.P.

The overlying unit (288-240 cm) in core Er is dark grey, clayey silt and silt. There is no change in lithology, however the abundance of shallow-water snail shells (*Viviparus quadratus dispiralis*) within the unit (Zhang, 1999) suggest that the lake became shallower. The diatom assemblage is dominated by the epiphytic species *Melosira granulata* (Zhang, 1999), consistent with the decreased water depth. By extrapolation, this phase occurred ca 10,900-9180 yr B.P.

The overlying unit (240-210 cm) in core Er is dark grey, clayey silt and silt. Snail shells are less abundant and consist largely of *Margarya* sp. (Zhang, 1999). The occurrence of this species of snail suggests the lake became somewhat deeper. The diatom assemblage is characterised by a decrease in *Melosira granulata* and an increase in *Cyclotella* sp. (Zhang, 1999), consistent with this interpretation. This phase occurred ca 9180-8100 yr B.P.

The overlying unit (210-200 cm) in core Er is dark grey, clayey silt and silt, with only occasional snail shells. Although there is no change in lithology, the further decrease in the abundance of snail shells (Zhang et al., 1999) suggest a further increase in water depth. The diatom assemblage is characterised by abundant *Cyclotella* sp. (45-60%) and the sparse representation of the epiphytic diatom *Fragilaria pinnata* (ca 1%) (Zhang et al., 1999). This assemblage is consistent with an increase in water depth. A sample from the top of this unit is radiocarbon dated to 7754±45 yr B.P., suggesting this phase occurred ca 8100-7800 yr B.P.

The overlying unit (200-160 cm) in core Er is dark grey, clayey silt and silt with abundant snail shells (*Viviparus quadratus dispiralis*). Although there is no change in lithology, the increase in the abundance of snail shells suggests the lake became shallower. The diatom assemblage is characterised by a decrease in the planktonic *Cyclotella* sp. (25-40%) and an increase in the epiphytic *Fragilaria pinnata* (1-6%), consistent with this interpretation. This phase occurred ca 7800-6330 yr B.P. There is a peat layer dated to 6550±200 yr B.P. in one of the unnamed cores from the western bay. Since this core was taken from a shallower-water site, closer to the lake margin than Er, the presence of peat is consistent with the lake-level lowering inferred from this snail-rich layer in Er.

The overlying unit (160-102 cm) is dark grey, clayey silt and silt, with very occasional snail shells. This change in the abundance of snail shells indicates an increase in water depth. The increase of planktonic *Cyclotella* sp. (50-60%, with a single sample registering 35%) and the decrease of epiphytic *Fragilaria pinnata* (<1%) are consistent with increased water depth. Two samples from 144-148 cm and the top of this unit are radiocarbon dated to 5825±40 and 4473±40 yr B.P. respectively, suggesting this phase occurred ca 6330-4500 yr B.P.

The overlying unit (102-90 cm) is brown to greyish brown, silt and clayey silt with snail shells (*Viviparus quadratus dispiralis* (*Heade*)). The increase in the abundance of snail shells suggests the lake became shallower. The decrease in *Cyclotella* sp. (30-40%) and the increase in *Fragilaria pinnata* (1-3%) are consistent with this interpretation. Using the sedimentation rate (0.22 mm/yr) between the date from the base of this unit and the top of the core, which is assumed to be modern, this phase occurred ca 4500-4030 yr B.P. A peat from one of the unnamed cores from the western bay of the lake has been radiocarbon dated to 4590±140 yr B.P. Since this core was taken from a shallower-water site, closer to the lake margin than Er, the presence of peat is consistent with the lake-level lowering inferred from this snail-rich layer in Er.

The overlying unit (90-78 cm) is brown to greyish brown, silty clay, with the very occasional presence of *Margarya* sp. The change in lithology is probably not significant. However, the occasional presence of *Margarya* (and the absence of *Viviparus*) suggests that the water became deeper (Zhang, 1998). The increase of planktonic *Cyclotella* sp. (40-45%) and the decrease of epiphytic *Fragilaria pinnata* (<1%) are consistent with increased water depth. This phase occurred ca 4030-3490 yr B.P. This phase of higher lake levels coincides with peat deposition at high elevation (+36 m) to the west of the modern lake, dated to ca 3650 yr B.P., and interpreted as indicating wetter conditions within the basin.

The overlying unit (78-52 cm) is brown silty clay, with the very occasional presence of *Margarya* sp. Thus, there is no change in lithology or the snail assemblage. However, changes in diatom assemblage, and specifically the decrease in *Cyclotella* sp. (25-40%) and the increase in the epiphytic *Fragilaria pinnata* (3-8%), suggest the lake became shallower. This phase occurred ca 3490-2330 yr B.P.

The overlying unit (52-44 cm) is brown silty clay, with the very occasional presence of *Margarya* sp. Thus, there is no change in lithology or the snail assemblage. The increase in the planktonic diatom *Cyclotella* sp. (45-65%) and the decrease in the epiphytic diatom *Fragilaria pinnata* (<1%) suggest the lake became deeper. This phase occurred ca 2330-1970 yr B.P.

The overlying unit (44-28 cm) is brown silty clay, with the very occasional presence of *Margarya* sp. Thus, there is no change in lithology or the snail assemblage. The decrease in the planktonic diatom *Cyclotella* sp. (25-45%) and the increase in the epiphytic diatom *Fragilaria pinnata* (1-3%) suggest the lake became shallower. This phase occurred ca 1970-1250 yr B.P.

The overlying unit (28-18 cm) is brown silty clay, with the very occasional presence of *Margarya* sp. Thus, there is no change in lithology or the snail assemblage. The increase in the planktonic diatom *Cyclotella* sp. (45-55%) and sparsity of the epiphytic diatom *Fragilaria pinnata* (<1%) suggest the lake became deeper. This phase occurred ca 1250-800 yr B.P.

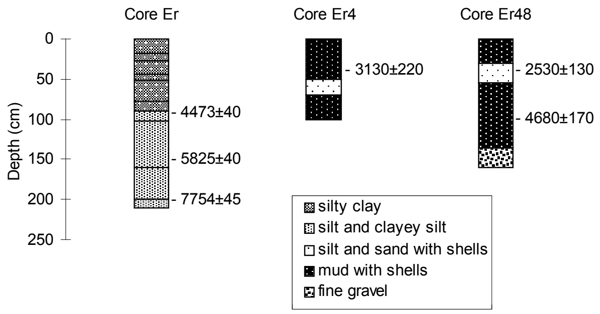
The uppermost unit (18-0 cm) is brown silty clay, with the very occasional presence of *Margarya* sp. Thus, there is no change in lithology or the snail assemblage. The decrease in the planktonic diatom *Cyclotella* sp. (ca 35%) and the increase in the epiphytic diatom *Fragilaria pinnata* (2-6%) suggest the lake became shallower. This phase occurred ca 800-0 yr B.P. There is no indication in the core sediments or the diatom assemblages to support the idea of increased. However, there is no change in lithology, or in the mollusc or diatom assemblages, showing that there is an increase in water depth at ca 470 yr B.P. indicated by the occurrence of snail deposits at elevations +2-3m above modern lake level ca 1500 away from the west coast of the lake.

Some additional information about changes in water depth during the late Holocene is provided by the records from core Er4 and core Er48.

The basal unit (100-70 cm) of core Er4, which was taken in a similar water depth to Er, is greyish brown mud with snail shells. The overlying unit (70-50 cm) is greyish brown medium to coarse sand with relatively abundant snail shells, suggesting a decrease in water depth. A single sample of snail shells from a depth of 60 cm within this unit was radiocarbon dated to 3130±220 yr B.P. The overlying unit (50-0 cm) is greyish brown mud with fewer snail shells, suggesting a return to deeper water. The interval of shallower conditions around ca 3130 yr B.P. is consistent with the record from the Er core.

The basal unit (160-135 cm) of core Er48 is mud with fine gravel. The overlying unit (135-55 cm) is grey mud with some snail shells. The finer texture of this unit indicates increased water depth. A sample of shells from 90-110 cm was radiocarbon dated to 4680±170 yr B.P. Using the sedimentation rate (0.021cm/yr) between this date and a date of 2530±130 yr B.P. (50-60cm) from an overlying unit, indicates that this mud was formed between 6400 and 2500 yr B.P. The overlying unit (55-30 cm) is silt and fine sand with snail shells (NIGL, 1989). The change in lithology indicates decreased water depth. A snail sample from the depth of 50-60 cm is radiocarbon dated to 2530±130 yr B.P., suggesting this phase of decreased water depth occurred ca 2530-1340 yr B.P. The overlying unit (30-0 cm) is grey mud, suggesting a return to somewhat deeper water after 1340 yr B.P. The occurrence of changes in in water depth during the later part of the Holocene is consistent with the interpretation of the Er record. However, the timing of the broad-scale changes from shallower to deeper to shallower to deeper are not consistent with the timing of changes in the Er core. We assume that this reflects the less-detailed examination of the core and extrapolation of the ages based on only two radiocarbon dates.

In the status coding, very low (1) is indicated by abundant shallow-water snail shells (*Viviparus*), with a diatom assemblage characterised by <40% *Cyclotella* and >1% *Fragilaria*, in the Er core; low (2) by moderately abundant deeper-water snail shells (*Margarya*) in the Er core; moderately low (3) by occasional deeper-water snail shells (*Margarya*), with a diatom assemblage characterised by <45% *Cyclotella* and >1% *Fragilaria*, in the Er core; intermediate (4) by occasional deeper-water snail shells (*Margarya*), with a diatom assemblage characterised by >45% *Cyclotella* and <1% *Fragilaria*, in the Er core; moderately high (5) by deposits containing no snail shells and with a diatom assemblage characterised by >45% *Cyclotella* in the Er core; high (6) by lacustrine deposits in core ZK14; very high (7) by lacustrine deposits in Cores ZK18 and ZK27; extremely high (8) by lacustrine deposits in Core ZK26.



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Radiocarbon dates

GC-635	34,090±3800	20.29-21.02 m, charcoal, ZK26
GC-628 (2)	23,050±1300	16.08-16.28 m, clay, ZK14
GC-830 (2)	22,265±1070	38.54-38.84 m, organic clay, ZK18
GC-628 (1)	21,650±830	4.64-4.94 m, organic clay, ZK14
GC-830 (1)	18,700±560	27.85-29.25 m, ZK18 (wrongly given as Z18 in Fig. 2.20 in NIGL et al., 1989)
GC-825 (2)	17,030±510	11.3-11.7 m, clay, ZK27(wrongly given as GC-826 (2) in Fig. 2.20 in NIGL et al., 1989)
GC-825 (1)	11,610±300	2.9-3.4 m, clay, ZK27 (wrongly given as GC-826 (1) in Fig. 2.20 in NIGL et al., 1989)
	7754±45	1.98-2.03 m, organic components, core Er
GC-1178	6550±200	ca 0.6-0.8 m, peat, core from the bay near the Shacun sand bar
	5825±40	1.44-1.48 m, organic components, core Er
	4680±170	0.9-1.1 m, snail shells, core 48
GC-632	4590±140	ca 0.5 m, peat, core from the bay near the Shacun sand bar
	4473±40	0.98-1.02 m, organic components, core Er
GC-630	3650±150	peat, +36 m above modern lake-level to the west of the lake
	3130±220	0.6 m, snail shells, core 4
	2530±130	0.5-0.6 m, snail shells, core 48
GC-1177	2230±90	snail shells, Shacun sand bar along the west coast
GC-634	1840±100	snail shells, Shacun sand bar along the west coast
GC-1072	1250±80	snail shells, Shacun sand bar along the west coast (wrongly given as GC-1172 in Fig. 2.21 in NIGL et al., 1989)
GC-631	470±150	snail shells, 1500 m away from the west coast of the modern lake

Coding

ca 34,090 yr B.P.	extremely high (8)
23,000-18,700 yr B.P.	high (6)
18,700-11,000 yr B.P.	very high (7)
11,680-10,900 yr B.P.	moderately high (5)
10,900-9180 yr B.P.	very low (1)
9180-8100 yr B.P.	low (2)
8100-7800 yr B.P.	intermediate (4)
7800-6330 yr B.P.	very low (1)
6330-4500 yr B.P.	intermediate (4)
4500-4030 yr B.P.	very low (1)
4030-3490 yr B.P.	intermediate (4)
3490-2330 yr B.P.	moderately low (3)
2330-1970 yr B.P.	intermediate (4)
1970-1250 yr B.P.	moderately low (3)
1250-800 yr B.P.	intermediate (4)
800-0 yr B.P.	moderately low (3)

Preliminary coding: February 1999
Second coding: May 1999
Third coding: October 2000
Final coding: 19-11-2000
Coded by BX, GY and SPH

3.41. Fuxian and Xingyun Lakes, Yunnan Province

Fuxian Lake (24.37-24.63°N, 102.81-102.95°E, 1720 m a.s.l.) and Xingyun Lake (24.33°N, 102.79°E, ca 1740 m a.s.l.) are two freshwater lakes in adjacent sub-basins within a large tectonic basin in the central Yunnan Plateau. Fuxian Lake lies in the Chengjiang Basin (area of 10,446km^2) and Xingyun Lake lies to the west in the Jiangchun Basin (325km^2). Both lakes overflow via small streams into the Nanpanjiang River and thence into the South China Sea. The Chengjiang and Jiangchun Basins are surrounded by high mountains (2500-2600 m a.s.l.) However, the watershed between the two lakes has an elevation of only ca 1740-1750 m a.s.l. (i.e. 10-20m higher than the modern lake level), and during the summer flood season there is an overflow from Xingyun Lake to Fuxian Lake (Zhu et al., 1989). Geomorphological and sedimentary evidence shows that the lakes were united into a single large lake during the early Holocene (Zhu et al., 1989; NIGL, 1990; Zhou et al., 1992). Thus, we consider the two lakes as a single basin. Fuxian Lake has a maximum water depth of 155m, a mean depth of 89.6m and an area of 211km^2. The modern sediment in the centre of Fuxian Lake in a water depth of ca 123m is brown clay (Zhu et al., 1989) and the ostracode assemblage (below 120m) is characterised by deepwater-benthic and planktonic species including *Fuxianhucythere inflata*, *Candona fuxianbuensis*, *Neochinocythere globra* and *Candona spinta* (NIGL, 1990). Xingyun Lake has a maximum water depth of 12m, a mean depth of 6.6m and an area of 35km^2. The lakes are supported by surface runoff from precipitation in the catchment (Zhu et al., 1989). The underlying bedrock in the catchment is mostly limestone. The regional climate is warm, with a mean annual temperature of 12-16°C, and wet, with ca 1000mm total annual precipitation (Zhou et al., 1992). The vegetation is subtropical evergreen forest (NIGL, 1990).

There are several sources of information about the evolution of the Fuxian-Xingyun lake basin. A record of the earliest phase of lacustrine deposition is provided by a profile through lacustrine terrace deposits from an unnamed opencut coal mine at 1820 m a.s.l. on the western slope of Jiangchuan Basin. This profile provides a record from ca 40,000-18,000 yr B.P. (Zhu et al., 1989). A profile (Nuimoucun Profile) through the lake terrace on the watershed between Fuxian and Xingyun Lakes, provides a record from the late glacial, ca 12,000 yr B.P. (Zhu et al., 1989). The top of this terrace has an elevation of 1745 m a.s.l. and the terrace is ca 500m from the shore of Lake Fuxian. Two profiles (Chengjiangxian Profile, Xiaolongcun Profile) from lake terraces ca 6-7 km north of Fuxian Lake also provide sedimentary records from the late glacial, ca 12,000-11,000 yr B.P. (Zhu et al., 1989). The highest terrace has an elevation of 1750m (Chengjiangxian Profile) and the second terrace has an elevation of 1740 m a.s.l. (Xiaolongcun Profile) A record of late Holocene sedimentation in the basin is provided by several cores from the lake-bottom sediments. Three short cores (Cores 80-36, 965, 80-16) (Zhu et al., 1989; Zhou et al., 1992) have been studied in detail. Core 80-36 is 1.6m long and was taken in a water depth of -75m; Core 965 is 1.8m long and was taken in a water depth of -123m; Core 80-16 is 1.85m long and was taken in a water depth of -150m. A further two cores (Core 972 from a water depth of -83m and Core 017 from a depth of -104m) provide information about the changing carbonate content of the sediments. Changes in water depth are reconstructed from shoreline evidence, and changes in lithology, carbonate content and ostracode assemblages in the cores. The interpretation follows the reconstructions in Zhu et al. (1989) and NIGL (1990).

There are 15 radiocarbon dates from the basin (Zhu et al., 1989; Zhou et al., 1993). Six of these dates are from the cores, 6 from the lacustrine terraces and 3 from archaeological sites within the basin. Two carbonate-rich samples (4635±110, 5530±150 yr B.P.) from Core 965 have reversed ages. Zhu et al. (1989) and Zhou et al. (1992) suggest that these samples are contaminated by old carbonate from the catchment and the ages are therefore too old. However, the other dates from the cores and the lacustrine terraces are believed to be reliable. The dates from the archaeological sites (2429±150, 3060±160 and 7030±240 yr B.P.) are on mollusc shells from midden deposits between 85 and 110m above modern lake level. Although these dates do not relate directly to the reconstructed lake-level changes, they do confirm that the lake was below ca 85m during the mid- to late-Holocene (Zhu et al., 1989).

The profile from the 1820m lacustrine terrace at the unnnamed coal mine site, bottoms out on Tertiary coal deposits. The overlying sediments consist of ca 2-3m in which lacustrine silt and clay alternates with three black peat units (Zhu et al., 1989; Zhou et al., 1992). Samples from the lower peat, the middle peat and the upper peat were radiocarbon-dated to > 40,000, 30,200±1500 and 19,478±500 yr B.P. respectively. Thus, the record from this profile indicates four deepwater phases occurring between sometime before 40,000 yr B.P. and 19,000 yr B.P., and three intervals (pre-40,000, 30,200 and 19,500 yr B.P.) when lake-level fell. The top of the lacustrine profile is at an elevation of 1820 m a.s.l. (+100m above modern lake level) and thus the lake must have been at least 100m higher than today when the lacustrine silts and clays were deposited. There is no indication of the water level when the peats were formed.

A 12m-high profile (Nuimoucun Profile) from the 1745m terrace on the watershed between the Fuxian and Xingyun Lakes bottoms out (below 11.5m) in pale-white gravels. This unit is probably colluvial in origin.The overlying sediment (11.5-10.2m) is lacustrine black clay, indicating that lake level was > 1735 m a.s.l. A sample from this unit was radiocarbon-dated to 11,831±415 yr B.P., suggesting this deepwater phase occurred before 11,800 yr B.P. The overlying sediment (10.2-8.4m) is grey-yellow sandy clay and fine sand, characteristic of shallow water deposition, suggesting decreased water depth after ca 11,800 yr B.P. The overlying sediment (8.4-7.0m) is coarse sand and gravel with crossbedding, characteristic of beach deposition, suggesting the lake level was ca 1738 m a.s.l. The uppermost sediment (above 7.0m) is gravel and sand, characteristic of fluvial-deltaic deposition. Thus the lake level must have been close to ca 1738 m a.s.l. when this unit was deposited. Unfortunately, there are no dates from this unit. However, the thickness of the deposit is consistent with deposition over at least a few hundred years.The cessation of sedimentation at this site indicates that lake level fell below ca 1738m sometime after ca 11,000 yr B.P.

The two profiles (Chengjiangxian Profile, Xiaolongcun Profile) from north of Fuxian Lake show a similar stratigraphy to one another, and also to the Niumoucun Profile. The basal deposits are colluvial gravels. The overlying unit is black lacustrine clay. A wood sample from this unit in Chengjiangxian Profile was radiocarbon-dated to 12,200±300 and an organic fraction sample from Xiaolongcun Profile to 11,995±420 yr B.P. This suggests that the deepwater phase recorded as occurring before 11,800 yr B.P. in Niumoucun Profile started as early as 12,200 yr B.P. (i.e. occurred between 12,200 and 11,800 yr B.P.). Unfortunately, there is no information about the absolute depths of the black clay unit in the Chengjiangxian and Xiaolongcun Profiles, and so it is not possible to use the information from these profiles to make a more precise estimation of the lake-level elevation. On the basis of shoreline feature 6-7km beyond the present lake margin in the north, and 2-3km beyond the present lake margin in the southwest, the area of the lake at ca 12,200-11,800 yr B.P. has been estimated as ca 350km^2 (Zhu et al., 1989; NIGL, 1990).

There is no lithological record for the early and middle Holocene. The absence of sediments in the lacustrine terraces dating to this interval indicate lake level was below 1740 m a.s.l. The presence of mollusc shells in midden sites dated to between ca 7000 and 2400 yr B.P. (Zhu et al., 1989) indicates that there was a lake present in the basin.

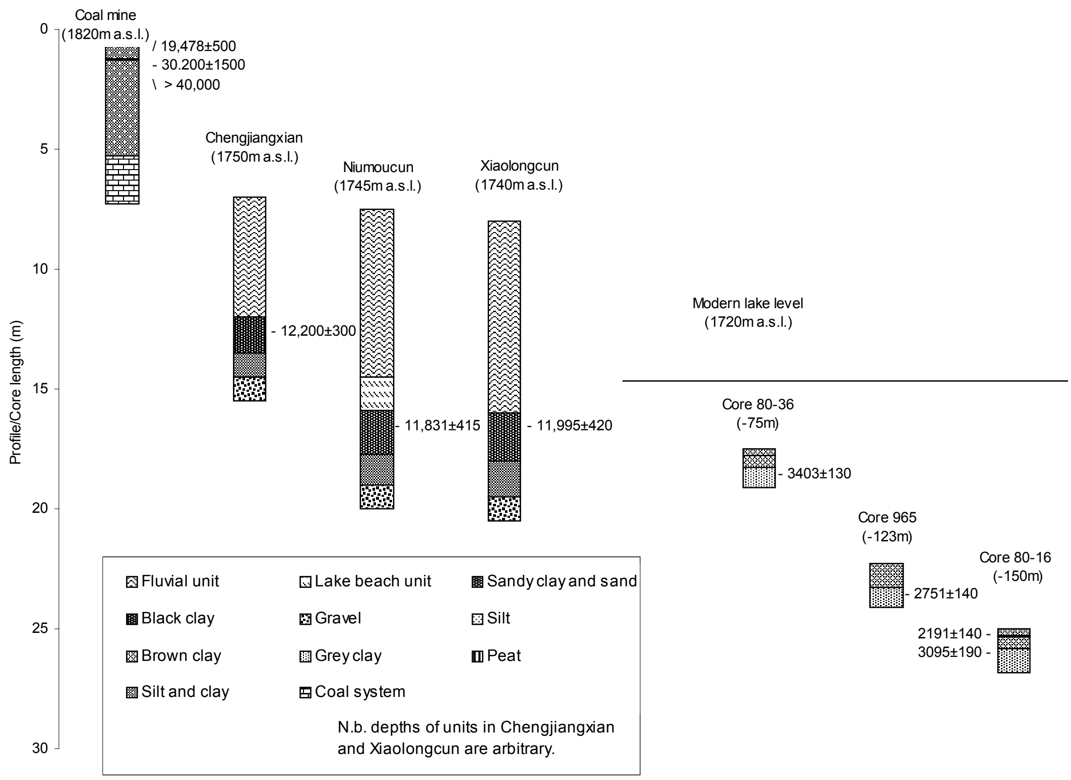
The lake-bottom cores provide a record of the late Holocene conditions in the basin. The sedimentary sequence from the cores is similar. The basal sediment is grey clay (0.85-1.85m in Core 80-16; 1.0-1.8m in Core 965; and 0.8-1.6m in Core 80-36), suggesting moderately deepwater conditions. However, the ostracode assemblage from Core 80-16 (-150m) does not contain the deepwater-benthic and planktonic species found in the deepest part (below -120m) of the lake today indicating that the water was at least 30m shallower than today (NIGL, 1990). The high values of Ca_2CO_3 (mean value 19.8%) from this grey clay unit in Core 972 and Core 017 indicates drier conditions in the catchment and greater carbonate inputs to the lake (NIGL, 1990), and thus is consistent with decreased water depth. Three samples from 0.85-1.85m in Core 80-16, 1.10-1.80m in Core 965 and 0.8-1.6m in Core 80-36 were radiocarbon-dated to 3095±190, 3403±130 and 2751±140 yr B.P yr B.P. respectively, indicating that this relatively shallow-water lacustrine phase occurred ca 3500-2700 yr B.P.

The overlying sediment is brown clay (0.85-0.35m in Core 80-16; 1.0-0m in Core 965; and 0.8-0.3m in Core 80-36). The change from grey to brown clay, which reflects decreased Ca_2CO_3 content (5%), indicates increased water depth. The sediments are similar to those being deposited today. The ostracode assemblage is characterised by the deepwater-benthic and planktonic species found in the lake today, including *Fuxianhucyfhere inflata*, *Candona fuxianbuensis*, *Neochinocyfhere globra* and *Candona spinta*, consistent with increased water depth. A sample from 0.35-0.85m in Core 80-16 was radiocarbon-dated to 2197±80 yr B.P., suggesting this deepwater phase occurred between 2700-420 yr B.P.

The overlying unit in all but the deepest-water core is silt (0.3-0.35m in Core 80-16 and 0.25-0.27m in Core 80-36). This change in lithology, and the fact that it is only registered in the shallower-water cores, indicates that water depth decreased slightly. This phase of shallowing occurred between 360-420 yr B.P. by extrapolation of sedimentation rate (0.083 cm/yr) between the radiocarbon dates on the underlying units.

The uppermost sediments in three cores all are brown clay (above 0.3m in Core 80-16 and above 0.25m in Core 80-36), indicating a return to deeper water conditions (similar to today) after ca 360 yr B.P.

In the status coding, very low (1) is lacustrine deposits with high carbonate contents and the absence of modern deepwater ostracode assemblages in the lake cores; low (1) by silt deposits in Cores 80-36 and 965; moderately low (3) by brown clay and an ostracode assemblage similar to modern in Cores 80-16, 80-36 and 965; intermediate (4) by nearshore deposits and fluvial-deltaic deposits at ca 1735 m a.s.l. in the lacustrine terraces at elevations between 1740 and 1750 m a.s.l.; moderately high (5) by lacustrine clay from 1745m terrace; high (6) peat deposits from the 1820m terrace; very high (7) by lacustrine silt and clay from 1820m terrace.



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Radiocarbon dates

	> 40,000	the lower peat, coal mine, 1820m terrace profile
	30,200±1500	the middle peat, coal mine, 1820m terrace profile
	19,478±500	the upper peat, coal mine, 1820m terrace profile
GC-599	12,200±300	black clay, Chengjiangxian, 1750m terrace profile
GC-597	11,995±420	wood, Xiaolongcun, 1740m terraces profile
GC-598	11,831±415	black clay, Nuimoucun, 1745m terrace profile
GC-705	5526±150	0.3-1.0m, brown clay, Core 965, ATO and not used (5530±150 given in Zhou et al., 1992)
GC-704	4635±110	0-0.3m, brown clay, Core 965, ATO and not used (4640±110 given in Zhou et al., 1992)
GC-600	3095±190	0.85-1.85m, grey clay, Core 80-16
GC-601	3403±130	1.10-1.80m, grey clay, Core 965 (3400±130 given in Zhou et al., 1992)
GC-706	2751±140	0.8-1.6m, grey clay, Core 80-36 (2750±140 given in Zhou et al., 1992)
GC-707	2197±80	0.35-0.85m, brown clay, Core 80-16 (2200±80 given in Zhou et al., 1992)

Coding

ca > 40,000 yr B.P.	high (6)
40,000-30,200 yr B.P.	very high (7)
ca 30,200 yr B.P.	high (6)
30,200-19,500 yr B.P.	very high (7)
ca 19,500 yr B.P.	high (6)
19,500-19,000 yr B.P.	very high (7)
19,000-12,200 yr B.P.	not coded
12,200-11,800 yr B.P.	moderately high (5)
11,800-11,000 yr B.P.	intermediate (4)
11,000-3400 yr B.P.	no record
3400-2700 yr B.P.	very low (1)
2700-420 yr B.P.	moderately low (3)
420-320 yr B.P.	low (2)
320-0 yr B.P.	moderately low (3)

Preliminary coding: 8-3-1999
Final coding: 27-3-1999
Coded by GY and SPH

3.42. Manxing Lake, Yunnan Province

Manxing lake (22°N, 100°36'E, ca 1160 m a.s.l.) is a small lake in the Menzhou Basin which lies in the southern part of the Yunnan Plateau, southwestern China. The Menzhou basin is a flat intermontane basin of tectonic origin. The Menzhou basin covers an area of ca 450 km^2. The highest mountain in this basin is over 2100 m a.s.l. The bedrock consist of granulites, shale and sandstone (Liu and Tang, 1987). There are several small lakes in the Menzhou basin, in addition to Manxing Lake. Manxing Lake has an area of ca 1.5 km^2 and a maximum water depth of 11 m (Tang, 1992). The lake is fed by small streams draining the catchment. The lake formerly drained to the Liusha River. However, the outlet has been recently dammed in order to use the lake water for irrigation (Tang, 1992). The regional climate is influenced by the southwest Asian monsoon (Indian monsoon) in summer and the east Asian winter monsoon (Siberia High Pressure cell) in winter. The mean annual temperature is 18.3^{o}C and the total precipitation is 1274 mm. The vegetation around the lake is evergreen forest. Subtropical evergreen forest occurs in the northern part of the basin, and tropical rainforest 30-60 km south of the basin (Tang, 1992).

A 5 m-long core (Core M), taken from the western part of the lake in a water depth of 3 m (Tang, 1992), provides a sediment record back to ca 27,400 yr B.P. Changes in water depth are reconstructed from changes in lithology and sedimentation rates. The chronology is based on seven radiocarbon dates from the core.

The basal unit (5.0-3.75 m) is black mud. The organic nature of these sediments (21-50%) suggests moderate water depths. The aquatic pollen assemblage is characterised by only Cyperaceae and *Polygonum*. Two samples from 5.0 and 4.5 m are radiocarbon dated to 27,360±850 and 26,280±590 yr B.P. respectively. Using the sedimentation rate calculatedg between these two dates (0.046 cm/yr), the upper boundary of the unit is dated to 24,660 yr B.P. Interpolation of the sedimentation rate (0.0178 cm/yr) between the uppermost date from the unit and a date of 20,650±440 yr B.P. from the overlying unit suggests the top of the unit dates to 22,050 yr B.P. Thus, this initial shallow water phase began ca 27,400 and ended sometime between 24,660 and 22,050 yr B.P.

The overlying unit (3.75-3.4 m) is black mud. A decrease in the organic content (13-21%) and a simultaneous increase in the abundance of quartz sand could indicate that the lake became slightly shallower. There is no significant change in the aquatic assemblage. A sample from 3.5 m is radiocarbon dated to 20,650±440 yr B.P. Using the sedimentation rate (0.0178 cm/yr) between the uppermost date from the underlying unit and this date, the upper boundary of the unit is dated to 20,090 yr B.P.

The overlying unit (3.4-3.3 m) is pure quartz sand. The marked change in lithology suggests that the lake became very shallow and may even have dried out. The unit is devoid of pollen, consistent with oxidation of the sediments under very shallow or dry conditions. A sample from a depth of 3.0m in the overlying unit has been radiocarbon dated to 11,870±380 yr B.P. Extrapolation of the sedimentation rate (0.01389 cm/yr) between this date and the overlying date would suggest that the top of the quartz sand unit dated to 14,030 yr B.P. The extremely low apparent sedimentation rates required to satisfy these discrepant estimates of the age of the upper and lower boundaries of the sand unit indicate that there must be a considerable sedimentary hiatus within the deposits. The occurrence of such an hiatus is consistent with drying out of the lake. We therefore suggest that the lake was low or dry from ca 20,000 to 14,000 yr B.P.

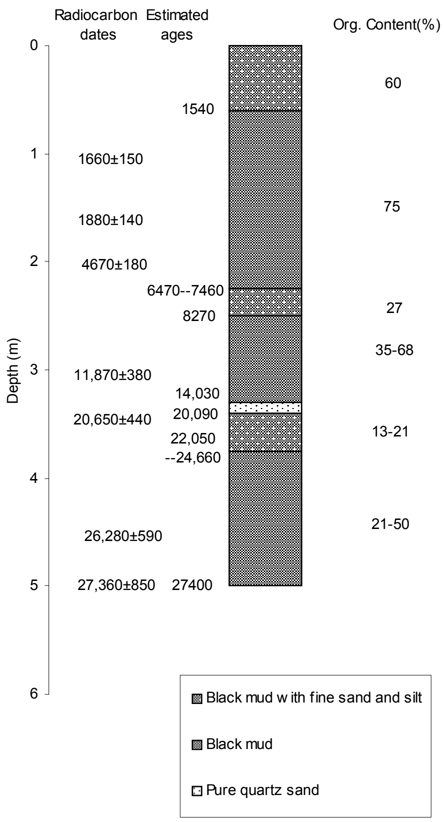
The overlying unit (3.3-2.5 m) is organic black mud, marking a return to lacustrine conditions after ca 14,000 yr B.P. The presence of aquatic pollen, including Cyperaceace and *Polygonum*, is consistent with this interpretation.

The overlying unit (2.5-2.25 m) is lacustrine black mud, with fine sand and silt. The increase in coarser material, coupled with a decrease in organic content (to ca 27% compared to ca 68% in the underlying unit) suggests that the lake became shallower. There is no significant change in the aquatic assemblage. Interpolation of the sedimentation rate between the bracketing radiocarbon dates indicates that this phase of somewhat shallower water occurred between 8270-7460 yr B.P. Extrapolation of the sedimentation rate between two radiocarbon dates from near the base of the overlying unit, however, suggests that the upper boundary is dated to 6470 yr B.P.

The overlying unit (2.25-0.6 m) is organic lake mud. The increase in organic content (75%) and the absence of sand- or silt-sized material is consistent with a return to somewhat deeper water conditions. There is no significant change in the aquatic assemblage. Three samples from 2.0 m, 1.75 m and 1.0 m are radiocarbon dated to 4670±180, 1880±140 and 1660±150 yr B.P. respectively. Extrapolation from these dates suggests this phase of somewhat deeper water conditions began between ca 7460 and 6470 yr B.P. and persisted until ca 1540 yr B.P.

The uppermost unit (0.6-0m) is organic black mud containing fine sand and silt. The increase in coarser-sized material and the decrease in organic content (60%) suggests the lake became shallower after 1540 yr B.P.

In the status coding, low (1) is indicated by coarse quartz sands and sedimentation rate evidence of a significant hiatus; intermediate (2) by moderately organic lake mud containing small amounts of sand and silt; high (3) by highly organic black mud without coarse material.



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Radiocarbon dates

M18	27,360±850	5.0 m, organic components
M16	26,280±590	4.5 m, organic components
M12	20,650±440	3.5 m, organic components
M10	11,870±380	3.0 m, organic components
M6	4670±180	2.0 m, organic components
M5	1880±140	1.75 m, organic components
M2	1660±150	1.0 m, organic components

Coding

27,400-22,050 yr B.P.	high (3)
24,660-20,090 yr B.P.	intermediate (2)
20,090-14,030 yr B.P.	low (1)
14,030-8270 yr B.P.	high (3)
8270 -6470 yr B.P.	intermediate (2)
7460 -1540 yr B.P.	high (3)
1540 -0 yr B.P.	intermediate (2)

Preliminary coding: January 1999
Second coding: January 1999
Final coding: 26-07-2000
Coded by BX, GY and SPH

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1. Appendix A: The Chinese Data Base
A1. CHDATA: LAKE INFORMATION

Basin	Province/ Region	Lat (°N)	Long (°E)	Elev (m)	Water type	Basin type	Origin	Geology	Basin area (km^2)	Lake area (km^2)	Mire (km^2)	Mean depth (m)	Maximum depth (m)	Basin precip. (mm/yr)	Basin evap. (mm/yr)	No.^{14}C dates	U-series	No. TL dates
Aiding Lake	Xinjiang	42.67	89.27	-155	salt	closed	faulting	unknown	150	5	n/a	n/a	0.5	5	1000-3000	4	0	0
Akesaiqin Lake	Xizang (Tibet)	35.20	79.83	4840	salt	closed	faulting	conglomerate, sandstone	n/a	160	n/a	n/a	12.6	20-40	2500	8	0	0
Aqigekule Lake	Xinjiang	37.07	88.37	4250	salt	closed	faulting	sandstone	n/a	345	n/a	9.8	n/a	100-300	n/a	3	0	0
Ashikule Lake	Xinjiang	35.73	81.57	4683	salt	closed	faulting and damming by lava	igneous	740	10.5	n/a	n/a	n/a	100-300	n/a	2	0	0
Baijian Lake	Gansu	39.15	104.17	1282	brackish	closed	tectonic	pre-Cambrian and Palaeozoic metamorphics and crystalline rocks, Mesozoic, Tertiary and Quaternary limestone and sandstone	>91000	n/a	small	n/a	1	115	2600	40 (1 ATY)	0	0
Baisuhai	Inner Mongolia	42.59	115.93	2000	fresh	closed	unknown	basalt	n/a	small	n/a	n/a	n/a	n/a	n/a	12	0	0
Balikun Lake	Xinjiang	43.70	92.80	1575	salt	closed	faulting	sandstone	10000	112.5	n/a	n/a	0.8-1	200-250	1000-1500	49 (only 28 used)	0	0
Bangge Lake	Xizang (Tibet)	31.75	89.57	4520	salt	closed	faulting	conglomerate, sandstone, argillite	n/a	135.4	n/a	0.7	1	308.3	2238.6	4	0	0
Bangongcuo	Xizang (Tibet)	33.70	79.42	4241	brackish	closed	faulting	unknown	n/a	412	n/a	18	n/a	60.4	2465.3	12	0	0
Beilikekule Lake	Xinjiang	36.72	89.05	4680	brackish	closed	faulting	sandstone	n/a	4.4	n/a	n/a	n/a	100-300	n/a	2	0	0
Big Ghost Lake	Taiwan	22.85	120.85	2150	fresh	closed	tectonic	miocene argillite and slate	0.9	0.1087	n/a	15.4	40	4200	n/a	5	0	0
Chaerhan Salt Lake	Qinghai	36.93	94.99	2675	salt	closed	faulting	metamorphics, sandstones, limestone, granite	n/a	460	5856 (playa)	0.2-1	1	28-40	3000	28	15	0
Chagannur	Inner Mongolia	43.27	112.90	920	salt	closed	faulting	sandstone and mudstone	2800	21	n/a	n/a	n/a	n/a	n/a	3	0	0
Chaiwopu Lake	Xinjiang	43.50	87.90	1092	salt	closed	faulting	Carboniferous-Permian detrital rock, with some carbonate and igneous rocks	400	29	n/a	4.22	6.1	50	3365.5	23 (2 not used)	8 (2 ATY, 1 not used)	0
Chitsai Lake	Taiwan	23.75	121.24	2890	fresh	closed	unknown	schist	n/a	0.28	n/a	n/a	n/a	n/a	n/a	4	0	0
Cuona Lake (CoNag Lake)	Xizang (Tibet)	32.03	91.47	4590	fresh	overflow	faulting	sandstone	1740-2610	174	n/a	n/a	>10	411.6	1770	4	0	0
Dachaidan-Xiaochaidan Salt Lakes	Qinghai	37.50	95.37	3110	salt	closed	faulting	Proterozoic metamorphics, Mesozoic sandstones and granite	n/a	91.92-104.9	390 (playa)	0.34	0.7	88.4	2080	10	0	0
Erhai	Yunnan	25.84	99.98	1974	fresh	overflow	tectonic	pre-Cambrian metamorphics, Ordovician sandstone, shale and carbonate	2565	249	n/a	10.2	20.7	n/a	n/a	20	0	0
Erjichuoer	Inner Mongolia	45.23	116.50	829.2	salt	closed	tectonic	sandstone, mudstone, conglomerate and metamorphics	700	10	16 (playa)	0.05	0.3	250-300	2000	2	0	0
Fuxian and Xingyun Lakes	Yunnan	24.48	102.87	1720-1740	fresh	overflow	tectonic	limestone	10771	211, 35	n/a	89.6, 6.6	155, 12	1000	n/a	12 (2 ATO)	0	0
Gounongcuo	Qinghai	34.35	92.20	4670	salt	closed	tectonic	unknown	n/a	2.9	n/a	n/a	n/a	200	2000	3	0	0
Hongshanhu Lake	Xizang (Tibet)	37.45	78.99	4870	salt	closed	faulting	sandstone	n/a	4.3	n/a	n/a	shallow	50	2500	4	0	0
Hulun Lake	Inner Mongolia	48.93	117.23	545	fresh	overflow	faulting	sandstone, mudstone and conglomerate with coal measures	n/a	2339	n/a	n/a	8	n/a	n/a	14 (1 ATO)	0	0
Jilantai	Inner Mongolia	39.75	105.70	1023	salt	closed	tectonic	metamorphics, sandstone and mudstone	2000	17.81	n/a	n/a	n/a	40-150	2800-4000	4	0	0
Longquanhu	Hubei	30.87	112.03	150	fresh	drained	unknown	sedimentaries	n/a	n/a	0.0005	n/a	n/a	900-1400	n/a	8 (1 not used)	0	0
Lop Basin	Xinjiang	40.29	90.80	780-795	salt	closed	tectonic	sandstone	300000	drained, 5350 (former lake)	n/a	n/a	n/a	20	2800-3000	10 (3 not used)	0	0
Manasi Lake	Xinjiang	45.45	86.00	251	salt	closed	faulting	granites, sedimentaries and limestone	11000	drained, 750 (fromer lake)	n/a	n/a	6 (former lake)	100	1000	11	0	0
Manxing Lake	Yunnan	22.00	100.60	1160	fresh	overflow	faulting	Granulites, shale, sandstone	450	1.5	n/a	n/a	11	1274	n/a	7	0	0
Nancun	Guangxi	24.75	110.42	160	fresh	drained	unknown	sandstone, shale, carbonate	n/a	drained	2.5	drained	drained	1560-2060	1255	8	0	0
Ningjingbo	Hebei	37.25	114.96	24 - 28	fresh	overflow	tectonic	sedimentaries	n/a	drained	n/a	drained	drained	500	n/a	5	0	0
North Tianshuihai Lake	Xizang (Tibet)	35.70	79.37	4797-4800	salt	closed	tectonic	slate	n/a	8	n/a	n/a	12.6	30	2500	8	0	0

Basin	Paleo-mag. Dating	Pb-210	Others	Record length (yr)	Lithology	Hiatus	Sed Rate	Aquatic pollen	Diatoms	Moll.	Ostr.	Geochem.	Geom.	Arch.	^{18}O or ^{13}C	Others	References	Other references	Coded by	Date
Aiding Lake	0	0	0	50000	yes	no	no	no	no	no	no	yes	no	no	no	no	Li et al., 1988	Yang et al., 1996	GY & SPH	Mar, 1999
Akesaiqin Lake	0	0	0	35000	yes	no	no	no	no	no	no	no	yes	no	no	aquatic plants	Li SJ et al., 1991; Li BY et al., 1991; Wang et al., 1990; Li et al., 1994	Wang et al., 1987; Hedin, 1922	GY & SPH	Mar, 1999
Aqigekule Lake	0	0	0	17000	yes	no	no	yes	no	no	no	no	yes	no	no	no	Li and Zhang, 1991; Li, 1992; Huang et al., 1996	Wang et al., 1987	GY & SPH	Dec, 1998
Ashikule Lake	0	0	0	16000	yes	no	no	yes	yes	no	no	yes	yes	no	no	no	Li and Zhang, 1991; Li, 1992; Huang et al., 1996		GY & SPH	Dec, 1998
Baijian Lake	0	0	0	39000	yes	no	no	no	no	yes	yes	no	yes	no	no	no	Pachur et al., 1995; Zhang and Wunnemann, 1995	Ma, 1998; Wunnemann and Pachur, unpublished manuscript	BX, GY & SPH	Jan, 2001
Baisuhai	0	0	0	13250	yes	no	no	no	no	yes	no	no	no	no	no	plant macros.	Cui and Kong, 1992; Cui et al., 1993		BX & SPH	Feb, 1999
Balikun Lake	0	0	0	37000	yes	no	no	no	no	no	no	yes	yes	no	no	no	Han and Dong, 1990; Han 1991; Han and Yuan, 1990; Han et al., 1993	Li, 1992; Xinjiang Geology Survey, 1983	GY & SPH	Jan, 2001
Bangge Lake	0	0	0	20100	yes	no	no	no	no	no	no	yes	yes	no	no	no	Zheng et al., 1989		GY & SPH	Jul, 2000
Bangongcuo	0	0	0	39600	yes	no	no	yes	no	no	yes	no	yes	no	no	no	Huang et al., 1989; Li YF et al., 1991, 1994; Huang et al., 1996; Li BY et al., 1991	Wang et al., 1987	GY & SPH	Jan, 2001
Beilikekule Lake	0	0	0	12740	yes	no	no	yes	no	no	no	no	yes	no	no	no	Li and Zhang, 1991; Li, 1992; Huang et al., 1996		GY & SPH	Jan, 1999
Big Ghost Lake	0	yes	0	2620	yes	no	no	no	no	no	no	yes	no	no	no	no	Luo et al., 1996; Luo and Chen, 1998	Luo, 1996	BX & SPH	Mar, 1999
Chaerhan Salt Lake	1	0	0	38000	yes	no	no	yes	no	yes	yes	yes	yes	no	no	no	Zheng et al., 1989; Huang et al., 1980; Huang and Chen, 1990; Du and Kong, 1983; Chen et al., 1990		GY & SPH	Nov, 2000
Chagannur	0	0	0	17850	yes	no	no	no	no	no	no	yes	no	no	no	no	Zheng et al.,1992; Xu, 1993		BX, GY & SPH	Mar, 1999
Chaiwopu Lake	1	0	0	30000	yes	no	no	yes	no	no	yes	no	yes	no	no	no	Gu et al., 1990; Huang, 1990; Li et al., 1990; Li and Yan, 1990; Wang and Jiao, 1989; Zhang and Wen, 1990		GY & SPH	Jan, 2001
Chitsai Lake	0	0	0	4700	yes	no	no	yes	no	no	no	no	no	no	no	no	Liew and Huang, 1994; Liew, 1999		BX & SPH	Feb, 1999
Cuona Lake (CoNag Lake)	0	0	0	35000	yes	no	no	no	no	yes	no	no	yes	no	no	no	Shen & Xu, 1993, 1994		GY & SPH	Jul, 2000
Dachaidan-Xiaochaidan Salt Lakes	0	0	0	24800	yes	no	no	no	no	no	no	yes	no	no	no	no	Zheng et al., 1989; Huang et al., 1980		GY & SPH	Nov, 2000
Erhai	0	0	0	34090	yes	no	no	no	yes	yes	no	no	yes	no	no	no	Zhang et al., 1998; Zhang, 1999; NIGL et al., 1989		BX, GY & SPH	Nov, 2000
Erjichuoer	0	0	0	15200	yes	no	no	no	no	no	no	yes	no	no	no	no	Zheng et al., 1992		BX, GY & SPH	Mar, 1999
Fuxian and Xingyun Lakes	0	0	0	40000	yes	no	no	no	no	no	yes	yes	yes	no	no	no	Zhu et al., 1989; Zhou et al., 1992; NIGL, 1990		GY & SPH	Mar, 1999
Gounongcuo	0	0	pollen	19600	yes	no	no	yes	no	no	yes	no	no	no	no	no	Li BY et al., 1994; Li YF et al., 1995; Li, 1996; Shan et al., 1996	Wang et al., 1987; Kong and Du, 1991; Shi et al., 1992	GY & SPH	Nov, 1999
Hongshanhu Lake	0	0	0	17200	yes	no	no	no	no	no	yes	no	no	no	no	aquatic plants	Li et al., 1994, 1995	Wang et al., 1987; Huang et al., 1985	GY & SPH	Mar, 1999
Hulun Lake	0	0	0	34000	yes	no	no	yes	yes	no	no	no	yes	no	no	no	Wang et al., 1995; Li, 1984		BX & SPH	Jan, 2001
Jilantai	0	0	0	17200	yes	no	no	no	no	no	no	yes	yes	no	no	no	Zheng et al.,1992; Geng and Cheng 1990		BX, GY & SPH	Mar, 1999
Longquanhu	0	0	0	9600	yes	no	no	yes	no	no	no	no	no	no	no	no	Li et al., 1992; Liu, 1993		BX & SPH	Feb, 1999
Lop Basin	0	0	0	20800	yes	no	no	yes	no	no	yes	yes	no	no	no	no	Zheng et al., 1991; Wu, 1994; Yan et al., 1983	Yan et al., 1998; Yang and Shao, 1993	BX, GY & SPH	Jan, 2001
Manasi Lake	0	0	0	32000	yes	no	no	yes	no	no	no	yes	no	no	yes	organic matter	Lin et al., 1996; Rhodes et al., 1996; Sun et al., 1994; Wei and Gasse, 1999	Zhang 1985; Huang et al. 1987; Gasse et al. 1987; Talbot and Livingstone 1989; Watts 1969; Wei and Gasse, 1999; XCET 1978	GY & SPH	Jul, 2000
Manxing Lake	0	0	0	27400	yes	yes	yes	no	no	no	no	no	no	no	no	no	Liu and Tang, 1987; Tang, 1992		BX, GY & SPH	Jul, 2000
Nancun	0	0	0	6400	yes	no	no	yes	no	no	no	no	no	no	no	no	Yao and Liang, 1993		BX, GY & SPH	Mar, 1999
Ningjingbo	0	0	pollen	18000	yes	no	no	yes	no	yes	yes	no	no	no	no	no	Guo and Shi, 1999; Guo, 2000		BX, GY & SPH	Mar, 1999
North Tianshuihai Lake	0	0	0	17750	yes	no	no	no	no	no	no	no	yes	no	no	aquatic plants	Li SJ et al., 1991; Wang et al., 1990		GY & SPH	Mar, 1999

Basin	Province/ Region	Lat (°N)	Long (°E)	Elev (m)	Water type	Basin type	Origin	Geology	Basin area (km^2)	Lake area (km^2)	Mire (km^2)	Mean depth (m)	Maximum depth (m)	Basin precip. (mm/yr)	Basin evap. (mm/yr)	No.^{14}C dates	U-series	No. TL dates
Salawusu Palaeolake	Shaanxi	37.70	108.60	1300	brackish	overflow	tectonic	Cretaceous sandstone	n/a	ca 100 (palaeolake)	n/a	n/a	n/a	250 - 450	n/a	10	0	6
Shayema Lake	Sichuan	28.83	102.20	2400	fresh	closed	faulting	unknown	n/a	0.04	n/a	10	n/a	800-1200	n/a	5	0	0
Toushe Lake	Taiwan	23.82	120.89	650	fresh	overflow	faulting	slate	n/a	n/a	0.1 (former lake)	n/a	n/a	2341	1098	16	0	0
Wulanwula Lake	Qinghai	34.80	90.50	4854	brackish	closed	faulting	igneous	n/a	544.5	n/a	n/a	>6.9	370	2000	5	0	0
Wulukekule Lake	Xinjiang	35.67	81.62	4687	salt	closed	faulting	volcanic	740	15	n/a	n/a	n/a	100-300	n/a	1	0	0
Xiaoshazi Lake	Xinjiang	36.97	90.73	4106	fresh	closed	faulting	sandstones	n/a	25	n/a	n/a	n/a	100-300	n/a	2	0	0
Xidadianzi	Jiling	42.33	126.37	614	peat bog	closed	volcanic-tectonic	volcanic	n/a	n/a	1.1	n/a	n/a	n/a	n/a	6 (1 ATY, 1 ATO)	0	0
Xingkai Lake (Khanka Lake)	Heilongjiang	45.17	132.17	69	fresh	overflow	tectonic	unknown	n/a	4800	n/a	6.2	10	n/a	n/a	1	0	3
Zabuye Lake	Xizang (Tibet)	31.35	84.07	4421	salt	closed	faulting	Cretaceous-Eocene acidic igneous, mudstone and sandstone	6680	243	n/a	0.7	2	192.6	2341.6	23 (1 ATO)	0	0
Zhacang Caka	Xizang (Tibet)	32.60	82.38	4328	salt	closed	faulting	Mesozoic sandstone and granite	n/a	109.75	n/a	n/a	2	151	2303	20 (13 not used)	0	0
Zigetangcuo	Xizang (Tibet)	32.08	90.83	4560	salt	closed	faulting	sandstone, mudstone	1840-2810	184	n/a	n/a	n/a	411.6	1770	2	0	0
Excluded lakes																		
Dahaizi Lake **	Sichuan	27.50	102.40	3660	fresh	closed	glacial	Unknown	n/a	0.15	n/a	n/a	n/a	n/a	n/a	3	0	0
Dajiu Lake *	Hubei	31.60	110.10	1700	fresh	overflow	unknown	metamorphic and sedimentary	16	n/a	10	n/a	n/a	n/a	n/a	6	0	0

Basin	Paleo-mag. Dating	Pb-210	Others	Record length (yr)	Lithology	Hiatus	Sed Rate	Aquatic pollen	Diatoms	Moll.	Ostr.	Geochem.	Geom.	Arch.	^{18}O or ^{13}C	Others	References	Other references	Coded by	Date
Salawusu Palaeolake	0	0	0	170000	yes	no	no	no	no	yes	no	yes	no	no	no	no	Yuan, 1978, 1988; Dong et al.,1983; Gao et al., 1985; Lu, 1985; Zheng, 1989; Su and Dong, 1994,1997; Sun et al., 1996	Teilhard de Chardin and Licent, 1924; CGS. 1956; Pei and Li, 1964; Qi 1975	GY & SPH	Mar, 1999
Shayema Lake	0	0	0	7700	yes	no	no	no	no	no	no	no	no	no	no	loss-on-ignition	Jarvis, 1993		SPH & BX	Feb, 1999
Toushe Lake	0	0	0	30000	yes	no	yes	no	no	no	no	no	no	no	no	loss-on-ignition	Huang et al., 1997; Liew et al., 1998		GY & SPH	Jul, 2000
Wulanwula Lake	0	0	0	18500	yes	no	no	yes	no	no	no	no	yes	no	no	no	Hu, 1995; Li YF et al., 1995; Shan et al., 1996	Kong and Du, 1991; Shi et al., 1992; Li BY, 1996	GY & SPH	Dec, 1998
Wulukekule Lake	0	0	0	7000	yes	no	no	no	no	no	no	no	yes	no	no	no	Li and Zhang, 1991; Li, 1992		GY & SPH	Dec, 1998
Xiaoshazi Lake	0	0	0	13000	yes	no	no	yes	yes	no	no	yes	yes	no	no	no	Li and Zhang, 1991; Li, 1992; Huang et al., 1996		GY & SPH	Mar, 1999
Xidadianzi Lake	0	0	0	10200	yes	no	no	yes	yes	no	no	no	no	no	no	no	Sun and Yuan, 1990		BX, GY & SPH	Mar, 1999
Xingkai Lake (Khanka Lake)	0	0	0	64000	yes	no	no	no	no	no	no	no	yes	no	no	no	Qiu et al., 1988	Wang et al, 1987	BX & SPH	Mar, 1999
Zabuye Lake	0	0	0	37600	yes	no	no	yes	no	no	yes	yes	yes	no	no	algae	Zheng et al., 1989, 1996; Qi and Zheng, 1995; Wu and Xiao, 1996		GY & SPH	Jan, 2001
Zhacang Caka	0	0	0	26000	yes	no	no	no	no	no	no	yes	yes	no	no	no	Huang et al., 1980; Zheng et al., 1989		GY & SPH	Nov, 2000
Zigetangcuo	0	0	0	19200	yes	no	no	no	no	no	no	no	yes	no	no	no	Shen & Xu, 1993, 1994		GY & SPH	Jul, 2000
Excluded lakes																				
Dahaizi Lake **	0	0	0	12400	yes	no	no	no	yes	no	no	no	no	no	no	no	Li and Liu, 1988			Feb, 1999
Dajiu Lake *	0	0	0	12000	no	yes	no	no	yes	no	no	no	no	no	no	no	Zhou and Li, 1993; Li et al., 1992			Feb, 1999

* Not sufficient information for lake status coding.
** Probably record of pond dynamics in peat bog. Not suitable for data base.
n/a not applicable or not available

A2. CHSTATUS: lake status coding

Basin	Lat (°N)	Long (°E)	Year (ka B.P.)																												
			0	0.5	1	1.5	2	2.5	3	3.5	4	4.5	5	5.5	6	6.5	7	7.5	8	8.5	9	9.5	10	10.5	11	11.5	12	12.5	13	13.5	14
Aiding Lake	42.67	89.27	1	1	1	2	2	4	4	4	4	4	4	3	3	3	4	4	4	4	5	5	3	3	3/5	5	5	5	5	6	6
Akesaiqin Lake	35.20	79.83	1	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	4
Aqigekule Lake	37.07	88.37	1	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	2	n/c	n/c	5	5	4	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Ashikule Lake	35.73	81.57	1	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	3	4	4	4	5	5
Baijian Lake	39.15	104.17	1	1	1	2/1	2	2	2	3/2	4/3	4/3	4/3	4	5/4/1	5/4/1	5/1	5/1	5	5	5	5/1	5/1	5/1	5/1	5/1	5/1	5/1	7/5	7	7
Baisuhai	42.59	115.93	1	1	1/4	4	4	4	4	3	3	3	3/2	3/2	2	2	2	2	2	3/2	3	3	3	3	3	3	3	3	3	n/c	n/c
Balikun Lake	43.70	92.80	1	1	1	1	1	1	1	1	1	9	9	9	9	9/1	9/1	9	9	2	2	3	3	3	3	3	1	7	7	7	7
Bangge Lake	31.75	89.57	1	2	2	2	2	2	2	2	2	2	2	2	2	3	2	2	4	5/2	4	4	4	5	5	5/3	3	3	3	3	3
Bangongcuo	33.70	79.42	1	1	2	3	3	3	5/2	5	5	5	5	5	5	1	4	4	4	4	4	7	7	3	4	4	4	4	4	4	7
Beilikekule Lake	36.72	89.05	1	1	1	1	1	1	1	2	2	2	2	4	4	4	4	1	1	1	1	1	1	1	1	1	3	3	1	n/c	n/c
Big Ghost Lake	22.85	120.85	2	2	1	1	1	2	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Chaerhan Salt Lake	36.93	94.99	1	1	1	1	1	1	1	1	3	3	2	2	2	2	2	2	2	4	4	2	2	2	2	2	2	2	2	2	2
Chagannur	43.27	112.90	7	7	7	7	7	8	8	3	1	5	2	3	3	3	3	4	4	1	3	3	3	3	2	2	2	2	2	3	2
Chaiwopu Lake	43.50	87.90	2	2/1	2/1	2/1	2/1	2/1	3	3	3/1	1	1	1	1	1	4	4	4	4	5	5	5	6/5	7	7	6	7	7	7	6
Chitsai Lake	23.75	121.24	2	2	2	2	1	1	1	1	1	3	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Cuona Lake (CoNag Lake)	32.03	91.47	1	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Dachaidan-Xiaochaidan Salt Lakes	37.50	95.37	1	1	1	1	4	4	4	4	3	3	3	3	3	3	3	7	7	7	7	7	2	2	2	2	2	6	6	6	6
Erhai	25.84	99.98	3	3	4	3	4	3	3	4	4	4/1	4	4	4	1	1	1	4	2	2	1	1	1	7/5	7/5	7	7	7	7	7
Erjichuoer	45.23	116.50	1	1	1	1	1	1	1	1	1	1	1	1	1	2	2	3	2	2	2	2	2	2	2	2	5	5	5	5	5
Fuxian and Xingyun Lakes	24.48	102.87	3	3	3	3	3	3	1	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	4	4	5	n/c	n/c	n/c	n/c
Gounongcuo	34.35	92.20	1	1	1	1	1	1	1	1	1	1	2/1	2	4	4	4	4	3	3	3	2	4	2	2	4/2	4	4	3/4	3	3
Hongshanhu Lake	37.45	78.99	1	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	3	3	2
Hulun Lake	48.93	117.23	1	1	1	1	1	1	1	1	1	1	1	1	3	3	3	1	1	1	1	1	1	4	5	7	7	7	n/c	n/c	n/c
Jilantai	39.75	105.70	1	1	1	1	1	1	1	1	1	3	3	2	2	2	2	4	4	4	4	4	3	3	3	3	3	3	3	3	3
Longquanhu	30.87	112.03	1	1	1	1	1	5	5	5	5	5	5	4	4	4	4	5	5	5	5	3	2	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Lop Basin	40.29	90.80	1	1	4	4	3	3	5	5/4	4	4	4	4	4	5/4	5	5	5	5	5	5/2	5/2	2	2	4	4	4	4	4	4
Manasi Lake	45.45	86.00	1	1/4	4	4	4	4	4	6	6	2	2	2	5	5	5	5	6	6	6	4	4	4	4	4	3	3	3	3	3
Manxing Lake	22.00	100.60	2	2	2	2	3	3	3	3	3	3	3	3	3	3/2	3/2	2	2	3	3	3	3	3	3	3	3	3	3	3	3
Nancun	24.75	110.42	3	3	3	1	1	1	1	1	1	1	1	1	1	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Ningjingbo	37.25	114.96	1	3	3	4	3	3	1	1	1	1	1	2	5	5	5	5	5	5	5	5	4	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
North Tianshuihai Lake	35.70	79.37	1	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	4	3	3	3	8/7	6/8	3
Salawusu Palaeolake	37.70	108.60	1	n/c	n/c	n/c	n/c	n/c	n/c	n/c	6	6	2/7	7	7	7	7	7	7	7	7	7	1	1	1	1	1	1	1	1	1
Shayema Lake	28.83	102.20	4	4	4	4	3	4	2	2	2	2	2	2	2	1	1	1	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Toushe Lake	23.82	120.89	n/c	n/c	n/c	3	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	3	1	1	1	1	1
Wulanwula Lake	34.80	90.50	1	1	1	1	2	2	2	2	4/2	4	4	4	4	4	4	4	3/4	3	3	3	3	3	3	n/c	n/c	2	n/c	n/c	n/c
Wulukekule Lake	35.67	81.62	1	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	2	2	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Xiaoshazi Lake	36.97	90.73	1	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	3	3	3	4	4	4	4	4	4	4	4	4	4	4	4	4	4	n/c	n/c
Xidadianzi	42.33	126.37	1	1	1	1	1	1	1	1	1	1	1	1	2	2	2	2	2	2	2	2	2	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Xingkai Lake (Khanka Lake)	45.17	132.17	1	1	1	1	1	1	1	1	1	1	1	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	n/c	n/c	n/c
Zabuye Lake	31.35	84.07	2	2	2	2	2	2	2	2	2	6	6	6	6	3	3	3	3	3	3	3	3	3	5/3	5/3	5/3	5/3	5	5	5
Zhacang Caka	32.60	82.38	1	1	2	2	2	2	2	2	2	3	3	4	4	4	6/4	6/4	6/2	6	6	6	6	6	6	6	6	6	6	6	6
Zigetangcuo	32.08	90.83	1	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c

Basin	14.5	15	15.5	16	16.5	17	17.5	18	18.5	19	19.5	20	20.5	21	21.5	22	22.5	23	23.5	24	24.5	25	25.5	26	26.5	27	27.5	28	28.5	29	29.5	30
Aiding Lake	5	5	5	6	6	6	5	5	5	5	5	5	5	5	5	5	5	5	5	5	5	7	7	7	7	7	7	7	7	7	7	7
Akesaiqin Lake	4	4	4	4	5	5	5	5	6	6	7	7	3	3	3	6	3	6	6	6	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Aqigekule Lake	n/c	n/c	n/c	n/c	2	3	2	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Ashikule Lake	5	5	5	1/6	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Baijian Lake	7	7	7	7/1	1	1	1	8/1	8	8/1	1	1	1	1	1	1	1	8/1	8	8	8	8	8	8	8	8	8	8	8	8	8	8
Baisuhai	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Balikun Lake	9	9	9	9	9	9	9	6	6	6	6	6	6	4	4	4	4	4	4	4	9	9	9	9	9	9	9	9	9	9	9	9
Bangge Lake	3	4	4	4	4	4	4	5	5	5	5	5	6	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Bangongcuo	6	6	6	3	n/c	n/c	n/c	n/c	10/8	10/8	10	10	10	10	10	10	8	9	9	9	9	9	9	9	9	9	9	8	8	8	10	10
Beilikekule Lake	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Big Ghost Lake	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Chaerhan Salt Lake	2	4	4	4/3	3	3	3	3	3	3	3	3	3	5	2	2	2	2	2	5	5	6	6	8	7	7	7	7	7	7	7	7
Chagannur	6	1	1	1	6/1	7	7	7	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Chaiwopu Lake	6	6	7	7	7	7	7	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	5
Chitsai Lake	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Cuona Lake (CoNag Lake)	n/c	n/c	2	6/2	6/2	6	6	6	4	6	4	4	6/4	3	2/3	2	2	5	5	2	2	7/2	7	7	7	7	7	7	7	7	7	7
Dachaidan-Xiaochaidan Salt Lakes	5	5	5	5	5	8	8	8	8	8	8	9	9	9	9	10	10	10	10	10	10	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Erhai	7	7	7	7	7	7	7	7	7	6	6	6	6	6	6	6	6	6	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Erjichuoer	4	4	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Fuxian and Xingyun Lakes	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	7	7/6	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7
Gounongcuo	3	3	3	5	5	5	4	4	4	4	4	0	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Hongshanhu Lake	2	2	6	6	6/4	5/4	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Hulun Lake	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	1	1	1	1	1	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	1	1	1	n/c	n/c
Jilantai	4	4	4	4	4	4	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Longquanhu	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Lop Basin	4	4	4	4	4	4	4	4	4	4	4	3	5	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Manasi Lake	3	3	3	3	3	3	3	3	3	3	3	3	3	3	3	3	3	3	3	3	3	3	3	3	3	3	3	3	3	3	3	3
Manxing Lake	1	1	1	1	1	1	1	1	1	1	1	1	2	2	2	2	2/3	2/3	2/3	2/3	2/3	3	3	3	3	3	n/c	n/c	n/c	n/c	n/c	n/c
Nancun	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Ningjingbo	n/c	n/c	n/c	n/c	n/c	n/c	n/c	1	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
North Tianshuihai Lake	3	3	3	3/2	3	3	6	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Salawusu Palaeolake	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	5	1	1	1	4/1
Shayema Lake	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Toushe Lake	1	1	1	1	1	1	1	3	3	1	1	1	3/1	3/1	3/1	3/1	3/1	3/1	3/1	3/1	3/1	3/1	1	1	1	1	1	1	1	1	1	1
Wulanwula Lake	n/c	n/c	n/c	n/c	n/c	n/c	n/c	5	5	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Wulukekule Lake	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Xiaoshazi Lake	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Xidadianzi	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Xingkai Lake (Khanka Lake)	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Zabuye Lake	5	5	5	5	5	5	7/5	7/5	7/5	7/5	7/5	7/5	7/5	7	7	7	7	7/4	7/4	4	4	4	4	4	4	4	4	4	4	6/4	6	6
Zhacang Caka	6	6	6	6	6	6	6	6	6	6	6	7/6	5	5	5	5	5	8/5	8	8	8	8	8	8	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Zigetangcuo	n/c	n/c	n/c	n/c	n/c	n/c	n/c	3	3	3	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c

A3. CHCOLSTA: lake status collapsed coding

Basin	Lat (°N)	Long (°E)	Year (ka B.P.)																												
			0	0.5	1	1.5	2	2.5	3	3.5	4	4.5	5	5.5	6	6.5	7	7.5	8	8.5	9	9.5	10	10.5	11	11.5	12	12.5	13	13.5	14
Aiding Lake	42.67	89.27	1	1	1	1	1	2	2	2	2	2	2	1	1	1	2	2	2	2	2	2	1	1	2/1	2	2	2	2	3	3
Akesaiqin Lake	35.20	79.83	1	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	2
Aqigekule Lake	37.07	88.37	1	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	1	n/c	n/c	3	3	2	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Ashikule Lake	35.73	81.57	1	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	1	2	2	2	3	3
Baijian Lake	39.15	104.17	1	1	1	1/2	2	2	2	2	2	2	2	2	2/1	2/1	2/1	2/1	2	2	2	2/1	2/1	2/1	2/1	2/1	2/1	2/1	2	2	2
Baisuhai	42.59	115.93	1	1	1/3	3	3	3	3	2	2	2	2/1	2/1	1	1	1	1	1	2/1	2	2	2	2	2	2	2	2	2	n/c	n/c
Balikun Lake	43.70	92.80	1	1	1	1	1	1	1	1	1	3	3	3	3	3/1	3/1	3	3	1	1	2	2	2	2	2	1	2	2	2	2
Bangge Lake	31.75	89.57	1	1	1	1	1	1	1	1	1	1	1	1	1	2	1	1	2	1/3	2	2	2	3	3	2/3	2	2	2	2	2
Bangongcuo	33.70	79.42	1	1	1	1	1	1	1/2	2	2	2	2	2	2	1	2	2	2	2	2	2	2	1	2	2	2	2	2	2	2
Beilikekule Lake	36.72	89.05	1	1	1	1	1	1	1	2	2	2	2	3	3	3	3	1	1	1	1	1	1	1	1	1	3	3	1	n/c	n/c
Big Ghost Lake	22.85	120.85	3	3	1	1	1	3	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Chaerhan Salt Lake	36.93	94.99	1	1	1	1	1	1	1	1	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2
Chagannur	43.27	112.90	3	3	3	3	3	3	3	2	1	2	1	2	2	2	2	2	2	1	2	2	2	2	1	1	1	1	1	2	1
Chaiwopu Lake	43.50	87.90	1	1/2	1/2	1/2	1/2	1/2	2	2	1/2	1	1	1	1	1	2	2	2	2	2	2	2	2	3	3	2	3	3	3	2
Chitsai Lake	23.75	121.24	2	2	2	2	1	1	1	1	1	3	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Cuona Lake (CoNag Lake)	32.03	91.47	1	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Dachaidan-Xiaochaidan Salt Lakes	37.50	95.37	1	1	1	1	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	1	1	1	1	1	2	2	2	2
Erhai	25.84	99.98	1	1	2	1	2	1	1	2	2	2/1	2	2	2	1	1	1	2	1	1	1	1	1	2/3	2/3	3	3	3	3	3
Erjichuoer	45.23	116.50	1	1	1	1	1	1	1	1	1	1	1	1	1	2	2	2	2	2	2	2	2	2	2	2	3	3	3	3	3
Fuxian and Xingyun Lakes	24.48	102.87	1	1	1	1	1	1	1	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	2	2	2	n/c	n/c	n/c	n/c
Gounongcuo	34.35	92.20	1	1	1	1	1	1	1	1	1	1	1/2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2
Hongshanhu Lake	37.45	78.99	1	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	2	2	1
Hulun Lake	48.93	117.23	1	1	1	1	1	1	1	1	1	1	1	1	2	2	2	1	1	1	1	1	1	2	2	3	3	3	n/c	n/c	n/c
Jilantai	39.75	105.70	1	1	1	1	1	1	1	1	1	2	2	2	2	2	2	3	3	3	3	3	2	2	2	2	2	2	2	2	2
Longquanhu	30.87	112.03	1	1	1	1	1	3	3	3	3	3	3	2	2	2	2	3	3	3	3	2	2	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Lop Basin	40.29	90.80	1	1	2	2	1	1	3	3/2	2	2	2	2	2	3/2	3	3	3	3	3	3/1	3/1	1	1	2	2	2	2	2	2
Manasi Lake	45.45	86.00	1	1/3	3	3	3	3	3	3	3	1	1	1	3	3	3	3	3	3	3	3	3	3	3	3	2	2	2	2	2
Manxing Lake	22.00	100.60	2	2	2	2	3	3	3	3	3	3	3	3	3	2/3	2/3	2	2	3	3	3	3	3	3	3	3	3	3	3	3
Nancun	24.75	110.42	3	3	3	1	1	1	1	1	1	1	1	1	1	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Ningjingbo	37.25	114.96	1	2	2	2	2	2	1	1	1	1	1	2	3	3	3	3	3	3	3	3	2	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
North Tianshuihai Lake	35.70	79.37	1	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	3	2	2	2	3	3	2
Salawusu Palaeolake	37.70	108.60	1	n/c	n/c	n/c	n/c	n/c	n/c	n/c	2	2	2/3	3	3	3	3	3	3	3	3	3	1	1	1	1	1	1	1	1	1
Shayema Lake	28.83	102.20	3	3	3	3	2	3	2	2	2	2	2	2	2	1	1	1	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Toushe Lake	23.82	120.89	n/c	n/c	n/c	3	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	3	1	1	1	1	1
Wulanwula Lake	34.80	90.50	1	1	1	1	2	2	2	2	2/3	3	3	3	3	3	3	3	2/3	2	2	2	2	2	2	n/c	n/c	2	n/c	n/c	n/c
Wulukekule Lake	35.67	81.62	1	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	3	3	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Xiaoshazi Lake	36.97	90.73	1	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	2	2	2	3	3	3	3	3	3	3	3	3	3	3	3	3	3	n/c	n/c
Xidadianzi	42.33	126.37	1	1	1	1	1	1	1	1	1	1	1	1	3	3	3	3	3	3	3	3	3	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Xingkai Lake (Khanka Lake)	45.17	132.17	1	1	1	1	1	1	1	1	1	1	1	3	3	3	3	3	3	3	3	3	3	3	3	3	3	3	n/c	n/c	n/c
Zabuye Lake	31.35	84.07	1	1	1	1	1	1	1	1	1	3	3	3	3	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2
Zhacang Caka	32.60	82.38	1	1	1	1	1	1	1	1	1	1	1	2	2	2	2	2	2/1	2	2	2	2	2	2	2	2	2	2	2	2
Zigetangcuo	32.08	90.83	1	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c

Basin	14.5	15	15.5	16	16.5	17	17.5	18	18.5	19	19.5	20	20.5	21	21.5	22	22.5	23	23.5	24	24.5	25	25.5	26	26.5	27	27.5	28	28.5	29	29.5	30
Aiding Lake	2	2	2	3	3	3	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	3	3	3	3	3	3	3	3	3	3	3
Akesaiqin Lake	2	2	2	2	2	2	2	2	3	3	3	3	1	1	1	3	1	3	3	3	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Aqigekule Lake	n/c	n/c	n/c	n/c	1	2	1	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Ashikule Lake	3	3	3	1/3	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Baijian Lake	2	2	2	2/1	1	1	1	3/1	3	3/1	1	1	1	1	1	1	1	3/1	3	3	3	3	3	3	3	3	3	3	3	3	3	3
Baisuhai	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Balikun Lake	3	3	3	3	3	3	3	2	2	2	2	2	2	2	2	2	2	2	2	2	3	3	3	3	3	3	3	3	3	3	3	3
Bangge Lake	2	2	2	2	2	2	2	3	3	3	3	3	3	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Bangongcuo	2	2	2	1	n/c	n/c	n/c	n/c	3/2	3/2	3	3	3	3	3	3	2	2	2	2	2	2	2	2	2	2	2	2	2	2	3	3
Beilikekule Lake	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Big Ghost Lake	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Chaerhan Salt Lake	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	3	3	3	3	3	3	3	3	3
Chagannur	2	1	1	1	1/2	3	3	3	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Chaiwopu Lake	2	2	3	3	3	3	3	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	2
Chitsai Lake	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Cuona Lake (CoNag Lake)	n/c	n/c	1	1/2	1/2	2	2	2	2	2	2	2	2	2	1/2	1	1	2	2	1	1	1/3	3	3	3	3	3	3	3	3	3	3
Dachaidan-Xiaochaidan Salt Lakes	2	2	2	2	2	2	2	2	2	2	2	3	3	3	3	3	3	3	3	3	3	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Erhai	3	3	3	3	3	3	3	3	3	2	2	2	2	2	2	2	2	2	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Erjichuoer	3	3	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Fuxian and Xingyun Lakes	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	3	3/2	3	3	3	3	3	3	3	3	3	3	3	3	3	3	3	3	3	3	3	3	3
Gounongcuo	2	2	2	3	3	3	2	2	2	2	2	1	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Hongshanhu Lake	1	1	3	3	2/3	2	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Hulun Lake	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	1	1	1	1	1	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	1	1	1	n/c	n/c
Jilantai	3	3	3	3	3	3	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Longquanhu	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Lop Basin	2	2	2	2	2	2	2	2	2	2	2	1	3	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Manasi Lake	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2	2
Manxing Lake	1	1	1	1	1	1	1	1	1	1	1	1	2	2	2	2	2/3	2/3	2/3	2/3	2/3	3	3	3	3	3	n/c	n/c	n/c	n/c	n/c	n/c
Nancun	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Ningjingbo	n/c	n/c	n/c	n/c	n/c	n/c	n/c	1	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
North Tianshuihai Lake	2	2	2	2	2	2	3	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Salawusu Palaeolake	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	2	1	1	1	1/2
Shayema Lake	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Toushe Lake	1	1	1	1	1	1	1	3	3	1	1	1	1/3	1/3	1/3	1/3	1/3	1/3	1/3	1/3	1/3	1/3	1	1	1	1	1	1	1	1	1	1
Wulanwula Lake	n/c	n/c	n/c	n/c	n/c	n/c	n/c	3	3	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Wulukekule Lake	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Xiaoshazi Lake	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Xidadianzi	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Xingkai Lake (Khanka Lake)	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Zabuye Lake	2	2	2	2	2	2	2/3	2/3	2/3	2/3	2/3	2/3	2/3	3	3	3	3	2/3	2/3	2	2	2	2	2	2	2	2	2	2	2/3	3	3
Zhacang Caka	2	2	2	2	2	2	2	2	2	2	2	2/3	2	2	2	2	2	2/3	3	3	3	3	3	3	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Zigetangcuo	n/c	n/c	n/c	n/c	n/c	n/c	n/c	3	3	3	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c

A4. CHDC: Dating control

Basin	Lat (°N)	Long (°E)	Year (ka B.P.)																												
			0	0.5	1	1.5	2	2.5	3	3.5	4	4.5	5	5.5	6	6.5	7	7.5	8	8.5	9	9.5	10	10.5	11	11.5	12	12.5	13	13.5	14
Aiding Lake	42.67	89.25	1D	2D	4D	5D	6D	7**	7**	7**	7**	7**	7**	7**	7**	7**	7**	7**	7**	7**	6D	5D	4D	2D	1D	3D	2C	2C	3C	3C	2C
Akesaiqin Lake	35.20	79.83	1D	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	1C
Aqigekule Lake	37.07	88.37	1D	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	2D	n/c	n/c	1D	2D	4D	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Ashikule Lake	35.73	81.57	1D	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	1D	2C	2C	2C	1C	2C
Baijian Lake	39.15	104.17	1D	2D	2D	1D	1D	3D	7*	7*	7*	7*	1D	1D	7*	7*	7*	7*	7*	7*	7*	7*	7*	7*	7*	7*	7*	7*	7*	4D	7
Baisuhai	42.59	115.93	1D	1C	1C	1D	1C	1C	1C	1D	1C	1C	1C	1C	1C	1C	1C	1C	1C	1C	1C	1C	2C	2C	2C	1C	2C	2C	1D	n/c	n/c
Balikun Lake	43.70	92.80	1D	2D	1D	2C	2C	1C	1C	2C	1D	2C	1C	1C	1C	1C	1C	1C	1C	1C	1C	1C	1C	1C	1C	1C	1C	1C	1C	1C	1C
Bangge Lake	31.75	89.57	1D	2C	2C	2C	2C	2C	2C	2C	2C	2C	2C	1C	1C	2C	1D	2C	1C	1C	1C	2C	1D	3D	4C	4C	4C	4C	3C	3C	4C
Bangongcuo	33.70	79.42	1D	2D	4D	7**	7**	7**	2D	1D	2C	2C	1C	2C	2C	1D	1D	7**	5D	3D	1D	2D	4D	4C	4C	4C	4C	3C	3C	4C	4C
Beilikekule Lake	36.72	89.05	1D	2D	4D	7	7	7	7	7	7	7	5D	4D	2D	1D	3D	4C	4C	3C	3C	3C	3C	4C	4C	4C	2D	1D	3D	n/c	n/c
Big Ghost Lake	22.85	120.85	1D	1C	1C	1C	1C	2D	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Chaerhan Salt Lake	36.93	94.99	1D	2D	2C	2C	1C	2C	2C	2C	1C	1C	1D	2C	2C	1C	2C	2C	1D	1C	1C	1D	3D	4C	4C	4C	3C	3C	3C	4C	4C
Chagannur	43.27	112.90	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	3D	1D	2C	1C	1C	1C	2C	1D	2C	2C	2C
Chaiwopu Lake	43.50	87.90	1D	2D	3D	1D	2D	1C	1C	1D	1D	2D	1C	1C	1C	1D	1C	1C	1C	1C	1C	1C	1C	1C	1C	1C	1D	2C	2C	1C	1C
Chitsai Lake	23.75	121.24	1D	1C	1C	1C	1C	1C	1C	1C	1C	1D	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Cuona Lake (CoNag Lake)	32.03	91.47	1D	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Dachaidan-Xiaochaidan Salt Lakes	37.50	95.37	1D	2D	4D	5D	4C	4C	4C	4C	3C	4C	4C	4C	4C	5D	3D	1D	2D	4C	4C	4C	4C	3C	3C	3C	4C	4C	4C	4C	2D
Erhai	25.84	99.98	1D	1D	1D	1D	1D	2D	3D	1D	2D	1C	1C	1C	1C	1C	1C	1C	1D	3D	5D	6D	6D	5D	3D	1D	2D	4C	4C	2C	3C
Erjichuoer	45.23	116.50	1D	2D	4D	5D	4C	4C	4C	4C	3C	4C	4C	4C	4C	5D	3D	1D	2D	2C	2C	2C	2C	2C	2C	2C	1D	2D	4D	5D	6D
Fuxian and Xingyun Lakes	24.48	102.84	1D	1C	1C	1C	1C	1C	1D	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	4D	2D	1D	n/c	n/c	n/c	n/c
Gounongcuo	34.35	92.20	1D	2D	4D	5D	6D	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	5D	3D	1D	1C	1C
Hongshanhu Lake	37.45	78.99	1D	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	3D	1D	1C
Hulun Lake	48.93	117.28	7	7	7	6D	5D	3D	1D	1C	1C	1C	1C	1C	1C	1C	1C	2C	2C	1C	1C	1C	1C	1C	1C	1C	1C	1C	n/c	n/c	n/c
Jilantai	39.75	105.70	1D	2D	4D	5D	6C	6C	6C	6C	4C	5C	5C	5C	4C	6C	6C	6C	6C	5D	3D	2D	1D	2C	2C	2C	2C	2C	2C	2C	2D
Longquanhu	30.87	112.03	1D	1C	1C	1C	1C	1C	1C	1C	1C	1D	1C	1C	1C	2C	1C	1C	1C	1C	1C	1D	3D	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Lop Basin	40.29	90.80	1D	2D	4D	5D	6D	5D	3D	1D	2D	1D	2C	2C	1C	2C	2C	1D	1D	2D	2D	1D	3D	5D	6C	4C	4C	4C	4C	4C	4C
Manasi Lake	45.45	86.00	1D	1C	1C	1C	1C	1C	1C	1C	1C	1D	2C	1C	1C	1C	2C	2C	2C	1C	1C	2C	1D	2D	4D	5D	6D	7	7	7	7
Manxing Lake	22.00	100.60	1D	1C	1C	1C	1D	2C	1C	1C	2C	1D	2D	4C	4C	4C	4C	4C	3C	3C	4C	4C	4C	4C	4D	2D	1D	3D	5D	6C	6C
Nancun	24.75	110.42	1D	1C	1C	1C	1C	1C	1C	1C	1C	1C	1C	1C	1C	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Ningjingbo	37.25	114.96	1D	1C	1C	1C	1C	1C	2C	1C	1C	1C	2C	1D	2C	2C	2C	3C	2C	2C	2C	1D	1D	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
North Tianshuihai Lake	35.70	79.37	1D	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	1D	2D	4D	3D	1D	2D	2D
Salawusu Palaeolake	37.70	108.60	1D	n/c	n/c	n/c	n/c	n/c	n/c	n/c	1C	1C	1C	2C	2C	2C	2C	2C	2C	2C	2C	1D	2D	4D	5D	6D	7	7	7	7	7
Shayema Lake	28.83	102.20	1D	1C	1C	1C	1C	1C	1D	1C	1C	1C	1C	1D	1C	1C	1C	1C	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Toushe Lake	23.82	120.89	n/c	n/c	n/c	2D	1C	1C	1C	1C	1C	1C	1C	1C	1C	1C	1C	1C	1C	1C	1C	1C	1C	1C	1C	1C	1C	1C	3D	4C	4C
Wulanwula Lake	34.80	90.50	1D	2D	4D	5D	6D	7**	7**	7**	7**	7**	7**	7**	7**	7**	7**	7**	7**	7**	7**	6D	5D	3D	1D	n/c	n/c	1D	n/c	n/c	n/c
Wulukekule Lake	35.67	81.62	1D	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	1D	2D	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Xiaoshazi Lake	36.97	90.73	1D	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	7	7	7	6D	5D	4D	2D	1D	1C	1C	1C	1D	2D	4D	5D	6D	7	n/c	n/c
Xidadianzi	42.33	126.37	1D	2C	2C	1C	1C	2C	1D	2C	1C	1C	1C	2C	1C	1C	1D	2C	1C	1C	2C	1D	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Xingkai Lake (Khanka Lake)	45.17	132.17	1D	7	7	7	7	7	7	6D	5D	4D	2D	1D	3D	5D	6D	7	7	7	7	7	7	6D	5D	3D	1D	2D	n/c	n/c	n/c
Zabuye Lake	31.35	84.07	1D	1C	1C	1C	1C	1C	1C	1C	1C	1C	1C	1C	1D	2D	1D	3D	3D	1D	2D	1D	2D	4D	5D	5D	3D	1D	2D	4D	5D
Zhacang Caka	32.60	82.38	1D	1C	1C	1D	2C	2C	1C	2C	2C	2C	1D	4D	6C	6C	6C	6C	4D	4D	1D	2D	4D	4C	4C	4C	3C	3C	2D	1C	1C
Zigetangcuo	32.08	90.83	1D	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c

Basin	14.5	15	15.5	16	16.5	17	17.5	18	18.5	19	19.5	20	20.5	21	21.5	22	22.5	23	23.5	24	24.5	25	25.5	26	26.5	27	27.5	28	28.5	29	29.5	30
Aiding Lake	2C	3D	1D	2D	4D	6C	6C	6C	6C	4C	4C	5C	5C	4C	4C	6C	6C	6C	6C	4D	2D	1D	3D	5D	6D	7	7	7	7	7	7	7
Akesaiqin Lake	1C	1C	1C	1D	1C	1C	1C	1C	1D	2C	2C	2C	2C	2C	2C	2C	2C	2D	4D	5D	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Aqigekule Lake	n/c	n/c	n/c	n/c	2D	1D	3D	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Ashikule Lake	2C	2C	1D	3D	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Baijian Lake	7	7	2D	3D	7	7	7	7	7	7	7	7	7	7	7	7	7	2D	1D	3D	5D	6D	5D	3D	1D	1D	2C	2C	1C	2C	2C	2C
Baisuhai	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Balikun Lake	1C	1C	1C	1C	1C	1C	1C	1D	2C	1C	1C	2C	1D	2C	2C	2C	1C	2C	2C	1D	1C	1C	1C	1C	1D	1C	1C	1C	1C	1C	1C	1C
Bangge Lake	4C	4C	4C	4D	2D	1D	3D	5D	6D	7	7	7	7	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Bangongcuo	4C	4C	3D	1D	n/c	n/c	n/c	n/c	2D	4D	5D	4C	4C	4C	4C	3C	4C	4C	4C	4C	5D	3D	1D	2D	2C	2C	2C	3C	2C	2C	2C	2D
Beilikekule Lake	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Big Ghost Lake	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Chaerhan Salt Lake	4C	3D	1D	1D	2D	2C	2C	2C	3C	2C	2C	2C	1D	2C	2C	2C	1C	2C	2C	2C	1C	1C	1C	1C	1D	3D	4C	2C	3C	3C	3C	3C
Chagannur	1C	2C	2C	2C	1D	3D	5D	6D	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Chaiwopu Lake	2C	1D	2D	4D	5D	6D	7	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	4D
Chitsai Lake	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Cuona Lake (CoNag Lake)	n/c	n/c	7**	7**	7**	7**	7**	7**	7**	6D	5D	4D	2D	1D	1C	2C	2C	2C	2C	2C	2C	2C	1D	3D	5D	6C	6C	6C	6C	4C	5C	5C
Dachaidan-Xiaochaidan Salt Lakes	1D	3D	4C	4C	4C	2C	2C	2C	4C	4C	4C	4C	2D	1D	2D	4D	5D	6D	7	7	7	7	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Erhai	3C	3C	2C	4C	3D	1D	2D	3D	1D	2C	2C	2C	1C	2D	1D	1C	1C	1C	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Erjichuoer	7	7	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Fuxian and Xingyun Lakes	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	2D	1D	3D	5D	6D	7	7	6C	6C	6C	6C	6C	5C	6C	6C	6C	6C	7	7	6D	5D	3D	1D
Gounongcuo	1C	1C	2C	2C	2C	2C	2C	2C	2C	2C	2D	4D	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Hongshanhu Lake	1C	1C	1C	1C	1C	1C	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Hulun Lake	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	7	7	7	7	7	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	7	7	7	n/c	n/c
Jilantai	4D	5D	6D	7	7	7	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Longquanhu	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Lop Basin	6C	6C	6C	5D	4D	2D	1D	3D	5D	6D	5D	4D	2D	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Manasi Lake	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7
Manxing Lake	6C	4C	4C	4C	4C	4C	4C	4C	4C	6C	5D	3D	1D	2D	4C	4C	2C	3C	3C	3C	2C	4C	4C	2D	1C	1C	n/c	n/c	n/c	n/c	n/c	n/c
Nancun	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Ningjingbo	n/c	n/c	n/c	n/c	n/c	n/c	n/c	7	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
North Tianshuihai Lake	1D	1D	1C	1C	1C	1C	1D	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Salawusu Palaeolake	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	7	6D	5D	4D	2D	1C	1C	1C	1C	1C
Shayema Lake	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Toushe Lake	3C	3C	3C	3C	4C	4C	3D	1D	2D	4D	5D	6C	6C	6C	6C	4C	5C	5C	5C	4C	6C	6C	6C	6C	5D	4D	2D	1D	1C	1C	1D	1D
Wulanwula Lake	n/c	n/c	n/c	n/c	n/c	n/c	n/c	1D	2D	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Wulukekule Lake	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Xiaoshazi Lake	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Xidadianzi	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Xingkai Lake (Khanka Lake)	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Zabuye Lake	6D	7	7	7	7	6D	5D	3D	1D	1C	1C	1C	2D	4D	5D	3D	1D	2D	2D	1D	4C	4C	3C	3C	4C	4C	4C	4C	4D	2D	1D	3D
Zhacang Caka	1C	1C	1D	2C	2C	2C	3C	2C	2C	2C	2C	1D	2D	4D	5D	6D	7	7	7	7	7	7	7	7	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c
Zigetangcuo	n/c	n/c	n/c	n/c	n/c	n/c	n/c	1C	1C	1C	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c	n/c

Note: * Dating control downgraded to 7 because of significant discrepancies between alternative chronologies.
** Dating control downgraded to 7 because of uncertainties about wheather coretop is modern.

A5. CHDATLST: Date list

Basin	Lat.	Long.	Number of dates	Lab no.	Date	Error +	Error -	Sample depth	Material	Core/preofile	Note
Aiding Lake	42.67	89.27	4								
					10900	420	420	4.95 m	clay and silt	core 86CK1	
					15700	300	300	7.81 m	clay and silt	core 86CK1	
					24900	1240	1240	11.6 m	clay and silt	core 86CK1	
					39700	4870	4870	14.5 m	clay and silt	core 86CK1	
Akesaiqin Lake	35.20	79.83	8								
					13920	400	400	0.1 m	silty clay	Profile N	
					15720	200	200	2.6 m	silty clay	Profile N	
					15960	240	240	2.8 m	silty clay	Profile N	
					16235	120	120	0.1 m	silt	Profile NE	
					18520	305	305	1.25 m	clay	Profile NE	
					22520	690	690	2.5 m	aquatic remains	Profile NE	
					33065	585	585	9 m	clay	Profile N	
					34735	820	820	10 m	clay	Profile N	
Aqigekule Lake	37.07	88.37	3								
					4705	108	108	0.1-0.2 m	silty clay	Profile East	
					6705	108	108	0.22-0.38 m	clay	Profile D	
					16765	149	149	0.15-0.2 m	aquatic plants	Profile B	
Ashikule Lake	35.73	81.57	2								
					11743	260	260	0.08-0.05 m	aquatic plants	Profile A	
					15256	100	100	2.0 m	clay	Profile A	
Baijian Lake	39.15	104.17	40								
				Hv18937	1405	60	60	top of section	carbonate	T6 terrace	
				Lu9318/N20	1910	60	60	bottom of section	carbonate	T6 terrace	
				Ld9303/N92	2561	85	85		peat	Baguamiaou	
				Hv19978	3315	130	130	0.57 m	carbonate	core B100	
				Lu932/N17	3660	55	55		snails	T5 terrace	
				Ld9302/N91	4645	120	120		peat	Baguamiaou	
				Lu933/N16	5250	70	70		snails	T4 terrace	
				Lu9322/N211	5510	60	60		carbonate	T4 terrace	
				Lu938/N31	5825	160	160		snails	Magang	
				Hv19979	6655	100	100	0.9 m	carbonate	core B100	
				Hv18933	7285	100	100		carbonate	Magang	
				Ld9301/N32	8211	115	115		organic carbon	Magang	
				Lu937/N21	8565	140	140		molluscs	Magang	
				Lu931	8720	105	105		shells	beach near Jiajiakeng	
				Lu9320/N43	10875	70	70		lake clay	Jianjiakeng profile	
				Lu9319/N41	12185	90	90		lake clay	Jianjiakeng profile	
				Lu9321/N42	12235	90	90		lake clay	Jianjiakeng profile	
				Lu/N 153	12817	140	140		carbonate	T2.2 terrace	
				Hv18936	16540	120	120		carbonate	T2.2 terrace	
				Hv19980	18620	325	325	2.9 m	carbonate	core B100	
				Lu9305/N221	22886	180	180		carbonate	1.8 m deep pit on playa	
				Lanzhou Univ.	23130	590	590		carbonate	T2.2 terrace	
				Hv19664	23370	380	380		shells	T2.1 terrace	
				Lanzhou Univ.	25920	900	900	ca 1.05 m	shells	Tudungcao	
				Ld9311/N114	26749	164	164	1.5-1.55 m	carbonate	Duantouliang	
				Hv19981	26900	1055	890	6.7 m	carbonate	core B100	
				Hv19982	27150	615	615	8.2 m	carbonate	core B100; ATY?	
				Lanzhou Univ.	27200	975	975		shells	T2.1 terrace	
				Lu936/N19	30330	560	560		shells	T2.1 terrace	
				Ld9310/N115	30360	175	175		shells	Duantouliang	
					31060	220	220	7.0 m		core B100	AMS
				Lu935/N18	32270	1236	1236		shells	above T2.1 terrace	
				Lu9315/N123	32435	840	840		shells	above T2.1 terrace	
				Lu9323/N111	33265	800	800		marl	Duantouliang	
				Lu934/N14	33500	1085	1085		shells	above T2.1 terrace	
				Lu9324/N112	35020	810	810	2.7-2.75 m	marl	Duantouliang	
					35660	420	420	9.3 m		core B100	AMS
				Lu9310/N102	36625	1630	1630	1.69-1.79 m	shells	Tudungcao	
					38650	970	970	3.1-3.2 m	carbonate	Duantouliang	
				Hv18934	38860	920	920	ca 2 m	shells	Tudungcao	
Baisuhai	42.59	115.93	12								
					1435	60	60	ca 5 cm	organic mud	core A	
					1695			ca 18 cm	mud	core B	
					3330			ca 44 cm	mud	core B	
					3640	70	70	ca 40 cm	organic mud	core A	
					4515	80	80	ca 80 cm	organic matter	core A	
					4680			ca 85 cm	organic matter	core B	
					6050			ca 117.5 cm	peat	core B	
					6625	60	60	ca 120 cm	peat	core A	
					7475			ca 162 cm	peat	core B	
					8270	70	70	ca 165 cm	peat	core A	
					9845	90	90	ca 210 cm	clay	core A	
					13020	60	60	ca 280 cm	clay	core A	
Balikun Lake	43.70	92.80	49								
					1218	80	80	0.02-0.36 m	mirabilite	core ZK00A	
					1700	65	65			accumulative barrier	

Basin	Lat.	Long.	Number of dates	Lab no.	Date	Error +	Error -	Sample depth	Material	Core/preofile	Note
					2310					core ZK00A	
					2370					core ZK00A	
					2640					core ZK00A	
					3270					core ZK00A	
					4130	116	116	0.78 m	mud	core ZK00A	
					5000			1.04 m	clay	core ZK00A	
					6618	89	89	1.2-1.25 m	mud	core ZK00A	
					7495	65	65	ca 2.9 m	mud	core ZK0024	
					8190					core ZK00A	
					8446	160	160	1.7 m	mirabilite clay	core ZK00A	
					8970					core ZK00A	
					9370	160	160	2.0 m	clayey mirabilite	core ZK00A	
					10084			2.2 m	clayey mirabilite	core ZK00A	
					10870					core ZK00A	
					12070	280	280	2.51 m	mirabilite clay	core ZK00A	
					12150	240	240	2.62 m	silty clay	core ZK00A	
					12530			2.84 m	silty clay	core ZK00A	
					12747					core ZK00A	
					13814					core ZK00A	
					13978					core ZK00A	
					14360	410	410	4.2 m	clay	core ZK00A	
					15170					core ZK00A	
					15829					core ZK00A	
					16176	360	360	5.0 m	silty mud	core ZK00A	
					16356	390	390	5.0 m	silty mud	core ZK00A	
					17800	470	470	5.6 m	clay	core ZK00A	
					20200					core ZK00A	
					20730	500	500	6.42 m	sandy clay	core ZK00A	
					21774					core ZK00A	
					23120					core ZK00A	
					24100			7.9 m	clay	core ZK00A	
					24310	225	225	ca 9.9 m	clay	core ZK0024	
					24489					core ZK00A	
					25670					core ZK00A	
					26413					core ZK00A	
					26618	670	670	8.9 m	silt and clay	core ZK00A	
					27350	600	600	10.3 m	clay	core ZK00A	
					28847					core ZK00A	
					29340			11.1 m	silty clay	core ZK00A	
					29470			11.5 m	silty clay	core ZK00A	
					30490			11.5 m	silty clay	core ZK00A	
					31432					core ZK00A	
					31950	110	110	12.1 m	silty clay	core ZK00A	
					32850	670	670	12.7 m	calcareous clay	core ZK00A	
					33710	170	170	13.0 m	calcareous clay	core ZK00A	
					35100	740	740	13.3 m	silty clay	core ZK00A	
					36700	829	829	13.6 m	clay	core ZK00A	
Bangge Lake	31.75	89.57	4								
					4425	90	90	1.22 m	clay	core CK2	
					7040	105	105	3.0 m	clay	core CK2	
					9795	115	115	8.55 m	marl	core CK2	
					16800	210	210	11.40 m	carbonate	core CK2	
Bangongcuo	33.70	79.42	12								
					3330	200	200			front of 15 m lake-delta terrace, Magazangbu	
					6750	235	235			top of 15 m lake-delta terrace, Magazangbu	
					9150	150	150	3.35 m	silt	Bangongcuo core	
					16100	220	220	12.3 m	clay	Bangongcuo core	
					18187	167	167	top of grass peat unit 4	plant remains	Taketuqiong profile II	
					25560	674	674	top of grass peat unit 3	plant remains	Taketuqiong profile II	
					29441	884	796	12.4 m	plant remains	Taketuqiong profile I	
					30302	685	685	middle of grass peat unit 2	plant remains	Taketuqiong profile II	
					35612	865	781	11.6 m	plant remains	Taketuqiong profile I	
					36454	847	847	middle of grass peat unit 2	plant remains	Taketuqiong profile II	
					39453	3263	3263	2.25 m	plant remains	Taketuqiong profile I	
					40602	3320	3320	bottom of grass peat unit 1	plant remains	Taketuqiong profile II	
Beilikekule Lake	36.72	89.05	2								
					6311	77	77	1.6 m	clay	Profile C	
					12253	280	280	4.75 m	aquatic plants	Profile C	
Big Ghost Lake	22.85	120.85	5								
					930	64	64	47-48 cm	organic component	93-cm long core	AMS
					1120	64	64	52-53 cm	organic component	93-cm long core	AMS
					1290	66	66	56-57 cm	organic component	93-cm long core	AMS
					1660	76	76	70 cm	organic component	93-cm long core	AMS
					2200	82	82	83 cm	organic component	93-cm long core	AMS
Chaerhan Salt Lake	36.93	94.99	43								
					3800			5.67 m	salt-bearing clay	core CK2022	
					4940			7.31 m	salt-bearing clay	core CK2022	

Basin	Lat.	Long.	Number of dates	Lab no.	Date	Error +	Error -	Sample depth	Material	Core/preofile	Note
					8120			12.03 m	salt-bearing clay	core CK2022	
					8850	210	210	1.30-1.70 m	salt-bearing muddy clay	core CK826	
					9170	100	100	13.38-13.78 m	salt-bearing clay	core CK2022	
					9300			13.87 m	salt-bearing clay	core CK2022	
					9310	310	310	2.5-2.9 m	salt-bearing muddy clay	core1308	
					13000	400	400	5.66-6.04 m	salt-bearing clay	core CK826	
					14900	100	100	14.89-15.19 m	salt-bearing clay	core 1308	
					15700	340	340	21.19-21.39 m	salt-bearing clay	core CK2022	
					16000			22.16 m	salt-bearing clay	core CK2022	
					18100	500	500	25.88-26.28 m	salt-bearing gypsum clay	core 1308	
					20600	410	410	35.52-34.87 m	salt-bearing clay	core CK2022	
					21200	210	210	27.02-27.42 m	salt-bearing clay	core 1308	
					21200	1050	1050	7.72-8.22 m	salt-bearing muddy clay	core CK826	
					24400	510	510	65.89-66.42 m	gypsum-bearing clay	core CK2022	
					24800	900	900	40.8 m	halite deposit	core CK-6	
					24800	470	470	68.14-68.54 m	gypsum-bearing clay	core CK2022	
					26400	700	700	84.5-85.0 m	clay	core CK2022	
					27600	1100	1100	16.3-16.7 m	salt-bearing clay	CK659	
					28200	900	900	52.3 m	halite deposit	core CK-6	
					28650	670	670	0.05-0.25 m	carbonate	shell-ridge	
					29700	500	500	44.17-44.52 m	clay	core1308	
					31800	2000	2000	96.36-96.76 m	clay	core CK2022	
					32200	1800	1800	54.9 m	halite deposit	core CK-6	
					33800	3000	3000	14.15-14.65 m	salt-bearing clay	core CK826	
					35100	900	900	1.1-1.2 m	carbonate	shell-ridge	
					38600	680	680	1.7-1.8 m	carbonate	shell-ridge	
					57500	8300	8300	57 m		CK-6	U/Th
					74800	9800	9800	92 m		CK-6	U/Th
					82300	10900	10900	99.2 m		CK-6	U/Th
					104000	9400	9400	151.9 m		CK-6	U/Th
					119500	11900	11900	169.6 m		CK-6	U/Th
					191900	24500	24500	206.7 m		CK-6	U/Th
					192900	34000	34000	249.7 m		CK-6	U/Th
					204000	38800	38800	260.9 m		CK-6	U/Th
					254200	45000	45000	283.4 m		CK-6	U/Th
					257600	60000	60000	323.0 m		CK-6	U/Th
					265000	43000	43000	271.2 m		CK-6	U/Th
					277700	73000	73000	297.7 m		CK-6	U/Th
					299000	92000	92000	354.0 m		CK-6	U/Th
					336,000	500,000-77,000	50077	392.2 m		CK-6	U/Th
					341,000	213,000-83,000	21383	378.2 m		CK-6	U/Th
Chagannur	43.27	112.90	3								
					9569	80	80	8.94 m	organic components	core 83-CK1	
					12554	80	80	10.3 m	organic components	core 83-CK1	
					16309	121	121	19.6 m	organic components	core 83-CK1	
Chaiwopu Lake	43.5	87.9	31								
				LB	1600	178	178	0.3 m	peat	T2 terrace profile (NE-a)	
				LB	2281	130	130	ca 1.5 m	clay	T1 terrace profile (NE)	
				LB	2834	59	59	ca 0.5 m	clay	T1 terrace profile (SW)	
				LB	3033	118	118	ca 1.2 m	clay	T1 terrace profile (NE)	
				LB-153	3696	90	90	0.35 m	clay	core CK1	
				LB	3740	60	60	0.5 m	peat	T2 terrace profile (NE-a)	
				LB	3830	70	70	ca 0.5 m	clay	T1 terrace profile (NE)	
				LB-154	3852	245	245	0.7-1.3 m	clay	core CK1	
				LB	4228	113	113	0.7 m	peat	T2 terrace profile (NE-a)	
				LB-156	5705	142	142	2.85-3.6 m	clay	core CKF	not used, large sample
				LB	6640	80	80	0.9 m	peat	T2 terrace profile (NE-a)	
				LB	6690	200	200	ca 3 m	peat	T2 terrace profile (NE-b)	
				LB	6958	217	217	1.1 m	mud	T2 terrace profile (NE-a)	
				LB	8180	120	120	1.4 m	mud	T2 terrace profile (NE-a)	
				LB	8320	80	80	ca 0.5 m	calcareous nodules	T3 terrace profile (NE)	
				LB	9470	80	80	1.9 m	clay	T2 terrace profile (NE-a)	
				LB	9650	130	130	ca 2 m	calcareous nodules	T3 terrace profile (NE)	
				GC-87086	9860	295	295	3.6 m	sand	core CKF	
				LB	10600	100	100	2.44 m	clay	T2 terrace profile (NE-a)	
				LB	12240	120	120	3 m	clay	T3 terrace profile (SW)	
				LB	14800	100	100	ca 5 m	clay	T2 terrace profile (NE-b)	
				LB	15030	140	140	4.3 m	clay	T3 terrace profile (SW)	
				LB-127	23821	766	766	16.0 m	clay	core CK1	not used, discrepant with U-series dates
					2500			1.25-1.35 m	carbonate	core CK1	U/TH, not used, discrepant with radiocarbon dates
					20000	2000	2000	14 m	carbonate	core CK1	U/Th
					31000	3000	3000	16 m	carbonate	core CK1	U/Th

Basin	Lat.	Long.	Number of dates	Lab no.	Date	Error +	Error -	Sample depth	Material	Core/preofile	Note
					62000	7000	7000	23-24 m	carbonate	core CK1	U/Th
					67000	8000	8000	53 m	carbonate	core CK1	U/Th
					211000	73000	73000	87-91 m	carbonate	core CK1	U/Th
					211000	90000	90000	57.7 m	carbonate	core CK1	U/Th
					303000	70000	70000	81-82 m	carbonate	core CK1	U/Th
Chitsai Lake	23.75	121.24	4								
				NTU-1732	1000	90	90	35 cm	wood	3.48 m core	
				NTU-1729	2650	60	60	170 cm	peat	3.48 m core	
				NTU-1733	3240	40	40	220-230 cm	peat	3.48 m core	
				NTU-1746	4300	50	50	312-317 cm	peat	3.48 m core	
Cuona Lake (CoNag Lake)	32.03	91.47	4								
					20916	1205	1205	4.4 m	sandy clay	core SG-89-2	
					21347	1130	1130	6.5 m	sandy clay	core SG-89-2	
					25397	964	964	7.6 m	black clay	core SG-89-2	
					35000			8.2 m	black clay	core SG-89-2	
Dachaidan-Xiaochaidan Salt Lakes	37.50	95.37	10								
					7630	140	140	2.94-3.39 m	salt-bearing clay	core CK3	
					9170	100	100	13.38-13.78 m	salt-bearing clay	core CK2022	
					14300	460	460	6.77-7.06 m	salt-bearing clay	core CK3	
					15700	340	340	21.19-21.39 m	salt-bearing clay	core CK2022	
					20600	410	410	34.52-34.87 m	salt-bearing clay	core CK2022	
					21000	1060	1060	9.4-9.84 m	clay	core CK3	
					24440	510	510	65.89-66.42 m	salt-bearing clay	core CK2022	
					24800	470	470	68.14-68.54 m	salt-bearing clay	core CK2022	
					26800	700	700	84.5-85 m	mud	core CK2022	
					31800	2000	2000	96.36-96.76 m	mud	core CK2022	
Erhai	25.84	99.98	20								
				GC-631	470	150	150		snail shells	1500 m from west coast	
				GC-1072	1250	80	80		snail shells	Shacun sand bar	
				GC-634	1840	100	100		snail shells	Shacun sand bar	
				GC-1177	2230	90	90		snail shells	Shacun sand bar	
					2530	130	130	0.5-0.6 m	snail shells	core 48	
					3130	220	220	0.6 m	snail shells	core 4	
				GC-630	3650	150	150		peat	36 m above modern lake-level west of lake	
					4473	40	40	0.98-1.02 m	organic components	core Er	
				GC-632	4590	140	140	ca 0.5 m	peat	core from bay near the Shacun sand bar	
					4680	170	170	0.9-1.1 m	snail shells	core 48	
					5825	40	40	1.44-1.48 m	organic components	core Er	
				GC-1178	6550	200	200	ca 0.6-0.8 m	peat	core from bay near the Shacun sand bar	
					7754	45	45	1.98-2.03 m	organic components	core Er	
				GC-825(1)	11610	300	300	2.8-3.4 m	clay	ZK27	
				GC-825(2)	17030	510	510	11.3-11.7 m	clay	ZK27	
				GC-830(1)	18700	560	560	27.85-29.25 m		ZK18	
				GC-628(1)	21650	830	830	4.64-4.94 m	organic clay	ZK14	
				GC-830(2)	22265	1070	1070	38.54-38.84 m	organic clay	ZK18	
				GC-628(2)	23050	1300	1300	16.08-16.28 m	clay	ZK14	
				GC-635	34090	3800	3800	20.29-21.02 m	charcoal	ZK26	
Erjichuoer	45.23	116.50	2								
					7503	864	864	ca 3.0 m	mud	83-CK1	
					12074	139	139	ca 7.8 m	clay	83-CK1	
Fuxian and Xingyun Lakes	24.48	102.87	12								
				GC-707	2197	80	80	0.35-0.85 m	clay	core 80-16	
				GC-706	2751	140	140	0.8-1.6 m	clay	core 80-36	
				GC-601	3403	130	130	1.1-1.8 m	clay	core 965	
				GC-600	3095	190	190	0.85-1.85 m	clay	core 80-16	
				GC-704	4635	110	110	0-0.3 m	clay	core 965,ATO	
				GC-705	5526	150	150	0.3-1.0 m	clay	core 965,ATO	
				GC-598	11831	415	415		clay	1745 m terrace, Niumoucun	
				GC-597	11995	420	420		wood	1740 m terrace, Xiaolongcun	
				GC-599	12200	300	300		clay	1750 m terrace, Chengjiangxian	
					19478	500	500	upper peat	peat	1820 m terrace	
					30200	1500	1500	middle peat	peat	1820 m terrace	
					>40000			lower peat	peat	1820 m terrace	
Gounongcuo	34.35	92.20	3								
					13035	155	155	315 cm	peat	core KX-1	
					15237	461	461	382 cm	peat	core KX-1	
					19210	480	480	530 cm	silty clay	core KX-1	
Hongshanhu Lake	37.45	78.99	4								
					13750	120	120	0.7 m	silty clay	6m-high section	
					15310	178	178	2.2 m	silty clay	6m-high section	
					16428	132	132	4.1 m	silty clay	6m-high section	
					17015	151	151	5.7 m	silty clay	6m-high section	
Hulun Lake	48.93	117.23	14								
				LT2	3080	80	80	0.34 m	palaeosol	Donglutian profile	

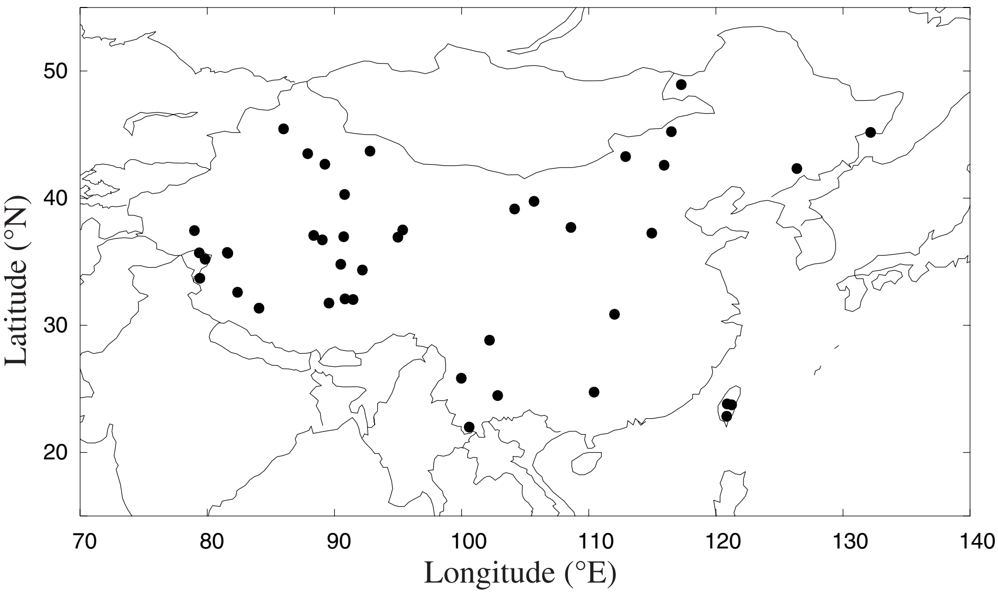
Basin	Lat.	Long.	Number of dates	Lab no.	Date	Error +	Error -	Sample depth	Material	Core/preofile	Note
				91HLH2	4790	100	100	ca 2.2 m	shell	Hangyun Station section	
				LT18	5270	80	80	2.1 m	clam shell	Donglutian profile	
				LT32	6710	200	200	3.6 m	peat	Donglutian profile	
				LT35	7070	200	200	3.9 m	clam shell	Donglutian profile	
				92HLD222	10280	285	285	6.9 m	organic clay	Donglutian profile	
				91HLD33	11300	225	225	9.47 m	organic clay	Donglutian profile	
				91HLB6	11410	210	210	ca 1.2 m	silty mud	Balongsabo profile	
				91HLD36	11750	550	550	9.75 m	organic clay	Donglutian profile	
				91HLD6	12700	230	230	12.15 m	organic clay	Donglutian profile	
				91HLG	19900	575	575	ca 1.2 m	peat	Gushan section	
				91HLD52	21000	625	625	ca. 6.7 m	charcoal	Donglutian profile	ATO?
				E8006	28900	1300	1300	bottom of Donglutian profile	paleo-tree stump	Donglutian profile	
				E8010	33760	1700	1700	bottom of Donglutian profile	coprolite of Mammuthus	Donglutian profile	
Jilantai	39.75	105.70	4								
					9782			10.4 m	organic components	core 83-CK1	
					9940	130	130		snail shells	beach deposits north of the modern lake	
					9959	130	130		snail shells	beach deposits east of the modern lake	
					13709			14.4 m	organic components	core 83-CK1	
Longquanhu	30.87	112.03	8								
				CG1976	2280	80	80	0.9-1.05 m	peat	LC2	
				CG1992	4410	180	180	1.9-2.05 m	clay	LC2	
				BK86075	6935	175	175	2.1-2.3 m	clay	LC1	
				CG1977	7000	140	140	2.9-3.05 m	clay	LC2	
				CG1993	8110	160	160	3.9-4.05 m	clay	LC2	
				CG1978	9155	195	195	4.9-5.05	silt	LC2	
				CG1994	9320	215	215	5.1-5.2 m	silt	LC2	
				CG1979	21910	200	200	5.6-5.72 m	silt	LC2	ATO
Lop Basin	40.29	90.80	10								
					3610	90	90	1.5 m	mud	L4	
					4725	150	145	0.5 m	clay	F4	
					7705	150	150	1.0 m	mud	F4	
					8000	165	160	1.3 m	mud	F4	
					9220	174	174	1.15 m	clay	K1	not used
					9360	120	120	3.1 m	mud	L4	
					17480	300	300	3.9 m	clay	F4	
					20780	300	300	8.5-8.83 m	mud	L4	
					23668	347	347	5.5 m	clay	K1	not used
					26172	479	479	ca 6.0 m	clay	K1	not used
Manasi Lake	45.45	86.00	11								
				H-603	330	80	80	1.52-1.54m	organics	LM I core	
				H-602	1140	50	50	1.713-1.728m	organics	LM II core	AMS
				H-601	1905	50	50	2.035-2.05m	organics	LM II core	AMS
				H-604	3440	120	120	2.82-2.84m	carbonate	LM II core	AMS
				H-690	4500	80	80	3.244-3.264m	carbonate	LM II core	AMS
					5310	95	95		carbonate	from exposed lacustrine sequence 20 km southwest of modern salt flat	
				H-605	7210	100	100	3.596-3.616m	carbonate	LM II core	AMS
				H-566	10030	560	560	4.005-4.015m	organics	LM II core	AMS not used for chronology (reversal)
				H-689	10120	100	100	3.979-3.999m	carbonate	LM II core	AMS
				H-644	32100	750	750	4.664-4.684m	carbonate	LM II core	AMS
				H-693	37800	1500	1500	4.684m	carbonate	LM III core	
Manxing Lake	22.00	100.60	7								
				M2	1660	150	150	1.0 m	organic components	core M	
				M5	1880	140	140	1.75 m	organic components	core M	
				M6	4670	180	180	2.0 m	organic components	core M	
				M10	11870	380	380	3.0 m	organic components	core M	
				M12	20650	440	440	3.5 m	organic components	core M	
				M16	26280	590	590	4.5 m	organic components	core M	
				M18	27360	850	850	5.0 m	organic components	core M	
Nancun	24.75	110.42	8								
				ZD2-343	235	42	42	0.05 m	clay	core 9	
				ZD2-342	1165	33	33	1.15 m	clay	core 9	
				GL-82006	1630	100	100	top of peat layer	peat	core unspecified	
				ZD2-344	3780	103	103	1.6 m	peat	core 4	
				ZD2-341	4735	108	108	2.1 m	peat	core 9	
				ZD2-345	5300	92	92	2.2 m	clay	core 4	
				ZD2-340	5875	74	74	2.45 m	clay	core 9	
				GL-82066	6400	115	115	bottom of peat layer	peat	core unspecified	

Basin	Lat.	Long.	Number of dates	Lab no.	Date	Error +	Error -	Sample depth	Material	Core/preofile	Note
Ningjingbo	37.25	114.96	5								
					1925	131	131	3.67 m	organic components	outcropping section	
					2642	96	96	4.0 m	organic components	outcropping section	
					5277	157	157	4.2 m	organic components	outcropping section	
					7050	110	110		organic material	Julu Profile	
					9750	350	350	5.6 m	organic components	outcropping section	
North Tianshuihai Lake	35.70	79.37	8								
					11235	650	650	3.8-4.0 m	silt	Profile T8	
					13070	200	200	1.5 m	sandy clay	Profile T50	
					14500	340	340	3 m	silty clay	Profile T5	
					15110	30	30	0.25 m	aquatic plant	Profile S9	
					15670	110	110	1.0 m	silty clay	Profile S9	
					17360	180	180	1.50-1.55 m	clay	Profile S9	
					17480	155	155	3.0 m	silt	Profile S10	
					17700	174	174	6.5 m	clay	Profile S10	
Salawusu Palaeolake	37.70	108.60	16								
					2300	90	90	ca 0.5m		Dishaogouwan Profile	
					3800	100	100	ca 1.5m		Dishaogouwan Profile	
					4700	100	100	ca 1.6m		Dishaogouwan Profile	
					5070	100	100	ca 1.8m		Dishaogouwan Profile	
					9500	100	100	ca 4.0-3.5m		Dishaogouwan Profile	
					9600	100	100	ca 4.0-3.5m		Dishaogouwan Profile	
					9700	120	120	ca 4.0-3.5m		Dishaogouwan Profile	
					27940	600	600	ca 10.5m		Dishaogouwan Profile	
					28170	1080	1080	ca 11.5m		Dishaogouwan Profile	
					30240	1280	1280	ca 13.60-13.65m		Dishaogouwan Profile	
					70900	6200	6200	23.0m		Dishaogouwan Profile	TL
					93000	14000	14000	37.5-38.0m		Dishaogouwan Profile	TL
					124900	15200	15200	43.0-43.5m		Dishaogouwan Profile	TL
					136000	15200	15200	44.0-44.5m		Dishaogouwan Profile	TL
					177000	14000	14000	44.5-45.0m		Dishaogouwan Profile	TL
					216000	22000	22000	58.0-58.5m		Dishaogouwan Profile	TL
Shayema Lake	28.83	102.20	5								
					1030	50	50	ca 2.35 m		core 1	
					2840	60	60	ca 4.45 m		core 1	
					5290	100	100	ca 6.2 m		core 1	
					7790	60	60	ca 7.85-7.8 m		core 1	
					10070	90	90	ca 10.5 m		core 1	
Toushe Lake	23.82	120.89	16								
					1840	50	50	0.3-0.4 m	peat	40m-long core	
					2230	50	50	0.8-0.9 m	peat	40m-long core	
					4230	50	50	1.72-1.82 m	peat	40m-long core	
					5640	60	60	3.1-3.2 m	peat	40m-long core	
					6480	60	60	4.2-4.3 m	peat	40m-long core	
					7370	60	60	4.73-4.83 m	peat	40m-long core	
					8270	70	70	5.35-5.45 m	peat	40m-long core	
					8780	60	60	6.1-6.2 m	peat	40m-long core	
					9600	130	130	7.05-7.07 m	peat	40m-long core	AMS
					9720	60	60	7.0-7.1 m	peat	40m-long core	
					10450	70	70	7.89-7.96 m	peat	40m-long core	
					12100	90	90	8.61-8.70 m	peat	40m-long core	
					12350	90	90	9.3-9.41 m	peat	40m-long core	
					18130	160	160	11.77-11.87 m	gyttja	40m-long core	
					28000	250	250	15.87-15.97 m	peat	40m-long core	
					29300	300	300	16.80-16.90 m	peat	40m-long core	
Wulanwula Lake	34.8	90.5	5								
					10997	252	252	ca 20 cm	lacustrine beach rock	Profile II.	
					11313	212	212	50-60 cm	mud	60cm-long core	
					11195	344	344	62 cm	plant remains	65cm-long core	
					12359	253	253	5-10 cm	muddy silt	Profile I	
					18217	390	390	ca 10 cm	lacustrine beach rock	Profile III.	
Wulukekule Lake	35.67	81.62	1								
					6505	77	77	0.4m	aquatic plant	Profile C	
Xiaoshazi Lake	36.97	90.73	2								
					8356	172	172	0.9 m	aquatic plants	Profile A	
					10693	238	238	1.5 m	aquatic plants	Profile A	
Xidadianzi	42.23	126.37	6								
					3070	95	95	3.25-3.0 m	peat	X1C	
					5830	160	160	6.25-6.0 m	silt and clay	X1C	
					6030	230	230	13.50-13.38 m	silt and clay	X1C	ATY
					6860	130	130	8.50-8.25 m	silt and clay	X1C	
					9970	160	160	13.25-13.0 m	silt and clay	X1C	
					12510	150	150	3.1-3.2 m	unknown	X3	ATO
Xingkai Lake (Khana Lake)	45.17	132.17	4								
					5430				archaeological material	on top of first sand ridge	
					12190	610	610		sand	2.01m below top of first (innermost) sand ridge	TL
					63000	3100	3100		sand	1.75m below surface of lake plain	TL
					63900	3190	3190		sand	1.15m below top of fifth (outermost) sand ridge	TL

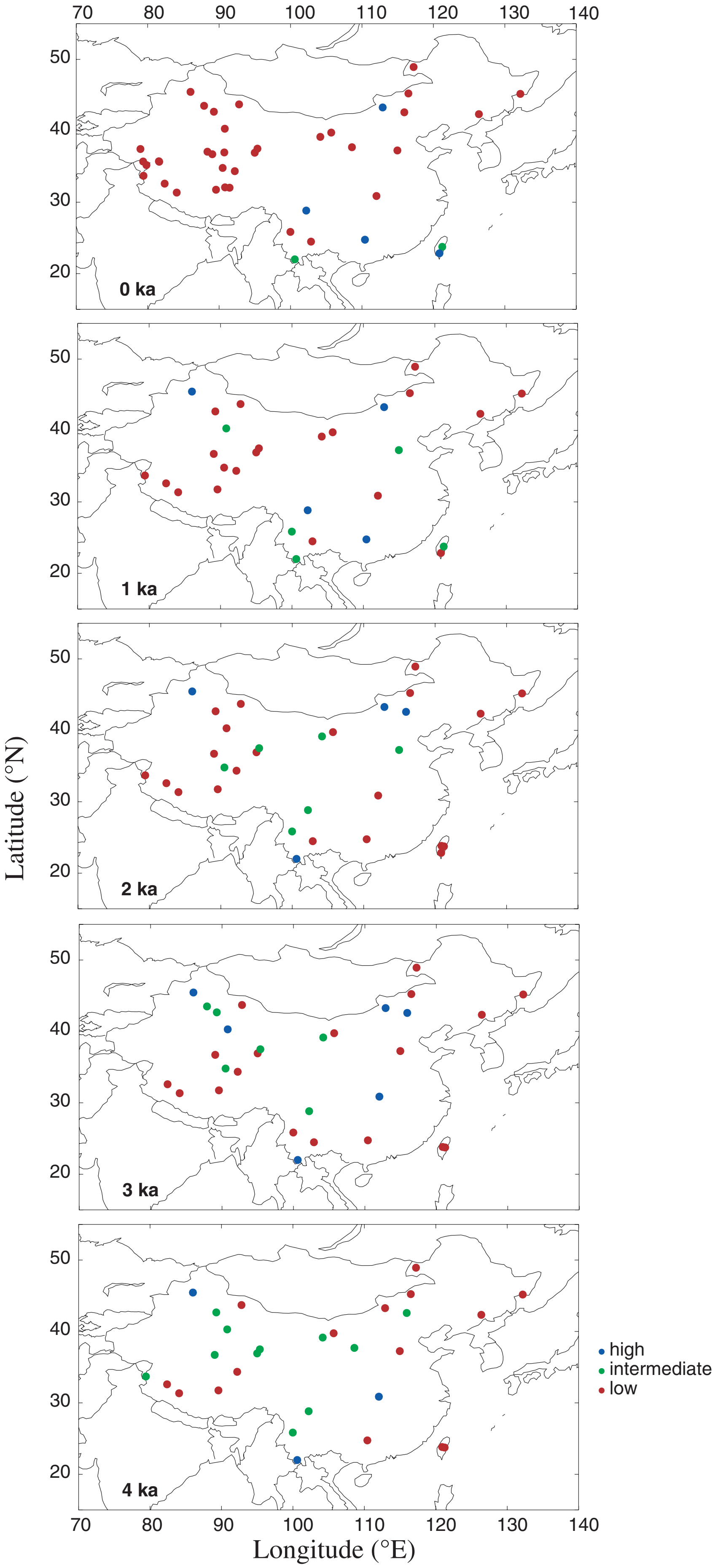
Basin	Lat.	Long.	Number of dates	Lab no.	Date	Error +	Error -	Sample depth	Material	Core/preofile	Note
Zabuye Lake	31.35	84.07	23								
					1350	70	70	ca 0.25 m	chlorate-mirabilite	core CK4	
					2170	150	150	ca 0.2 m	borax-carbonate	F32	
					3150	70	70	ca 0.8 m	carbonate	CK4	
					3150	70	70	ca 1.20 m	calcareous clay	CK8	
					3530	70	70	ca 0.4 m	calcareous clay	CK8	probably ATO
					3950	80	80	ca 2.0 m	mud	CK5	
					4190	160	160	ca 1.36 m	mirabilite clay	core ZK91-2	AMS
					4470	80	80	ca 1.6 m	calcareous clay	CK8	
					5315	135	135	4440 m a.s.l	carbonate	southern profile	
					5770	80	80	ca 2.1 m	mud	core CK4	
					5980	80	80	ca 4.5 m	mud	core CK4	
					5990	100	100	1.81-1.91 m	clay	core ZK91-2	AMS
					6840	170	170	ca 1.5 m	carbonate	core CK1	
					8725	135	135	4470 m a.s.l.	carbonate	Chaduixiong	
					9510	165	165	4440 m a.s.l.	carbonate	northern profile	
					12535	180	180	4485 m a.s.l.	calcareous clay	Jiadonglongba	
					18620	300	300	ca 3.1 m	carbonate	core CK2	
					20080	450	450	ca 5.5 m	carbonate	core CK2	
					22130	235	235	6.25-6.35 m	clay	core ZK91-2	AMS
					22610	500	500	ca 4.2 m	carbonate	core CK1	
					22670	380	380	4480 m a.s.l.	carbonate	Jiuer	
					23770	600	600	4510 m a.s.l.	calcareous clay	Chaduixiong	
					29330	420	420	12.54-12.64 m	calcareous clay	core ZK91-2	AMS
Zhacang Caka	32.60	82.38	20								
					1400	690	690	1.35-1.55 m	salt-rich clay	core 78CK3	
					3000	810	810	1.50-1.65 m	mud	76CK2	
					3840	130	130	2.47-1.70 m	mud	76CK11	
					4780	180	180	4.5-4.7 m	gypsum and halite-rich clay	core 78CK3	
					5600	150	150	1.74-1.94 m	mud	76CK1	
					5710	130	130	2.78-2.99 m	mirabilite-bearing mud	76CK7	
					6070	210	210	2.30-2.55 m	mud	76CK3	
					6200	160	160	3.90-4.10 m	mud	76CK5	
					6500	160	160	3.65-3.88 m	mud	76CK2	
					7000	110	110	4.80-5.00 m	mud	76CK4	
					7800	210	210	3.80-4.10 m	mirabilite-bearing mud	76CK3	
					8000	130	130	2.40-2.70 m	mud	76CK1	
					8090	130	130	5.45-5.70 m	mud	76CK11	
					9060	120	120	6.3-6.5 m	muddy clay	core 78CK2	
					10400	250	250	5.15-5.35 m	mud	76CK9	
					10900	200	200	4.65-4.85 m	mud	76CK1	
					13400	160	160	13.2-13.4 m	muddy clay	core 78CK3	
					15400	160	160	14.2-14.4 m	muddy clay	core 78CK3	
					15600	600	600	9.6-9.9 m	muddy clay	core 78CK2	
					20000	350	350	15.6-15.8 m	clay with sand and gravel	core 78CK3	
Zigetangcuo	32.08	90.83	2								
					17707	405	405	12.0m	silty clay	River Simaijiuqu profile	
					19220	387	387	14.5m	silty clay	River Simaijiuqu profile	

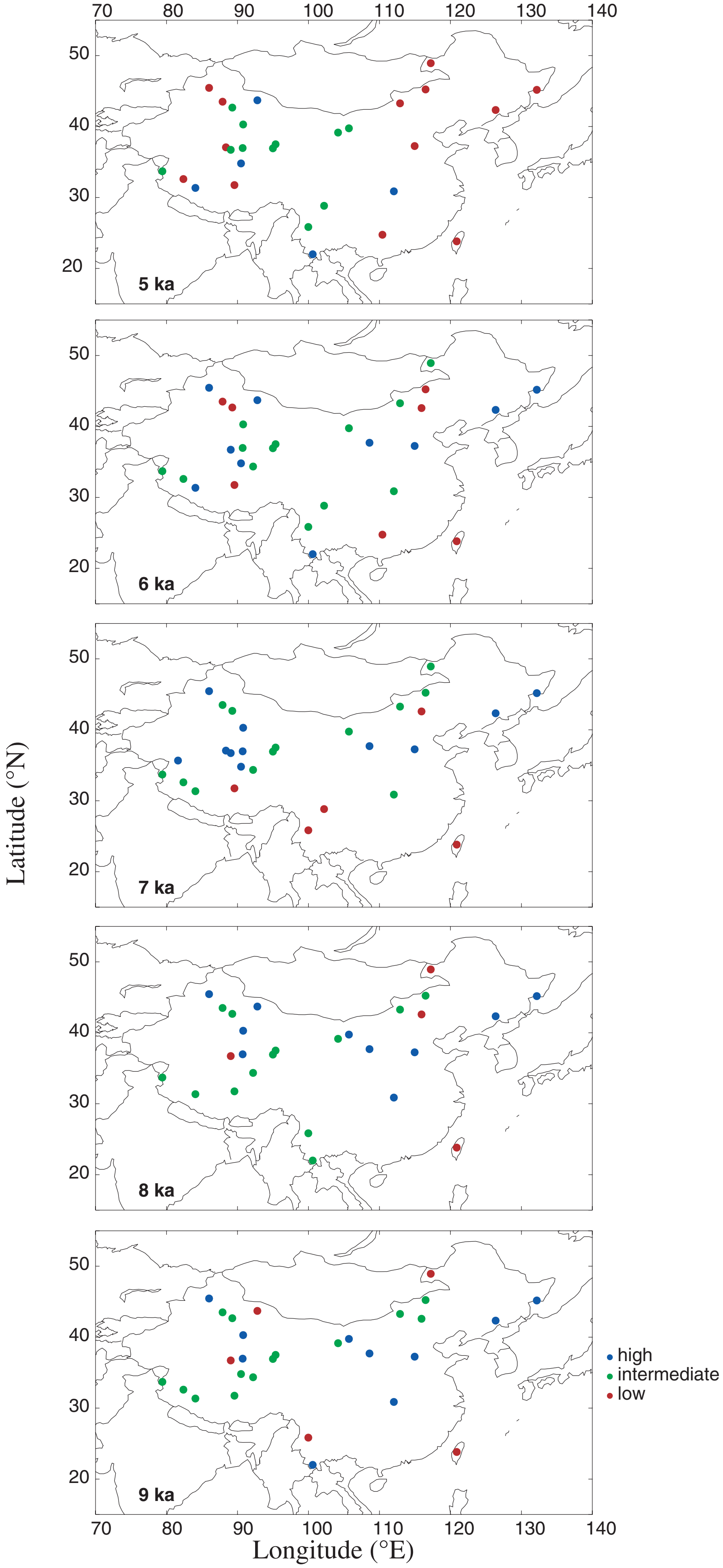
6. Appendix B: Maps of Lake Status During the Late Quaternary in China

B1. Site map



B2. Lake-Status maps from 18 to 0ka B.P.




70
80
90
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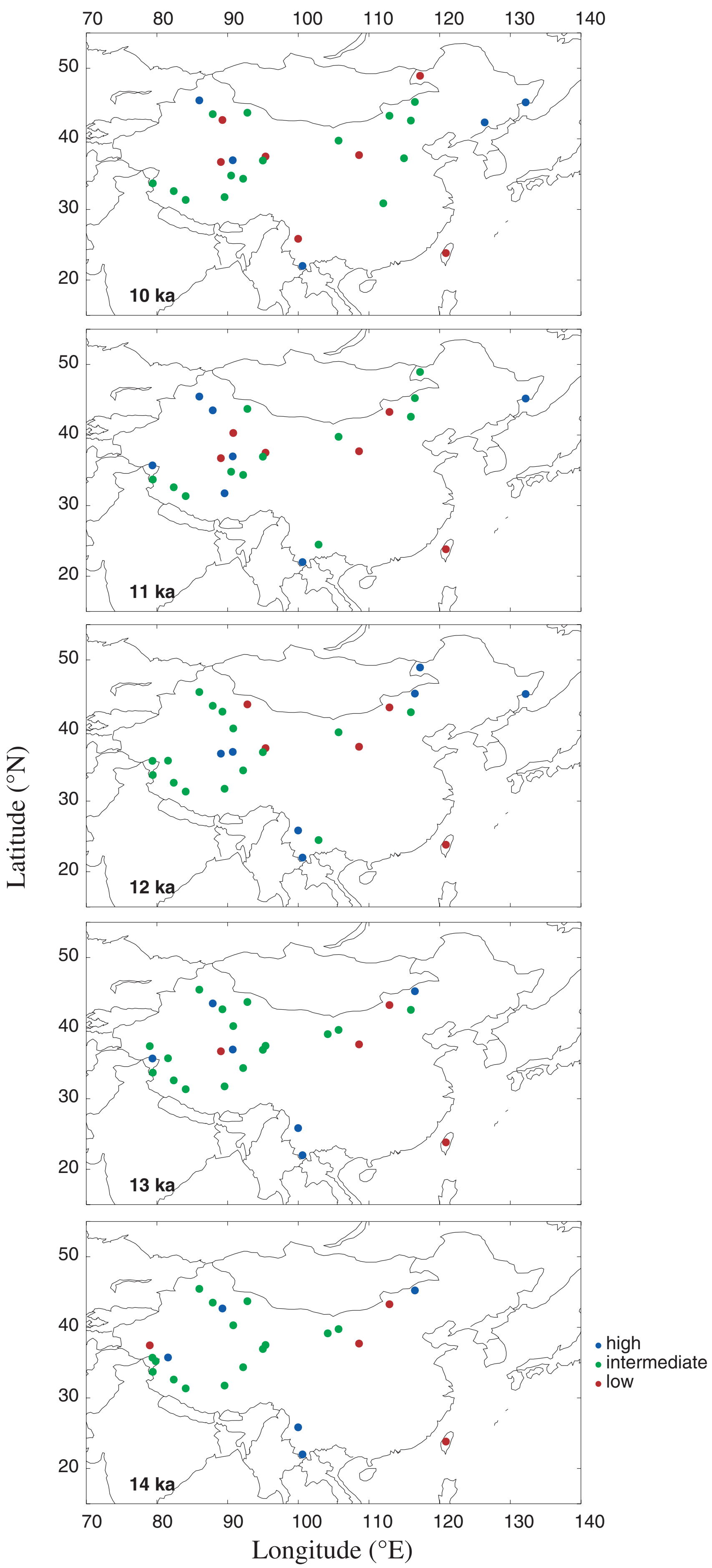
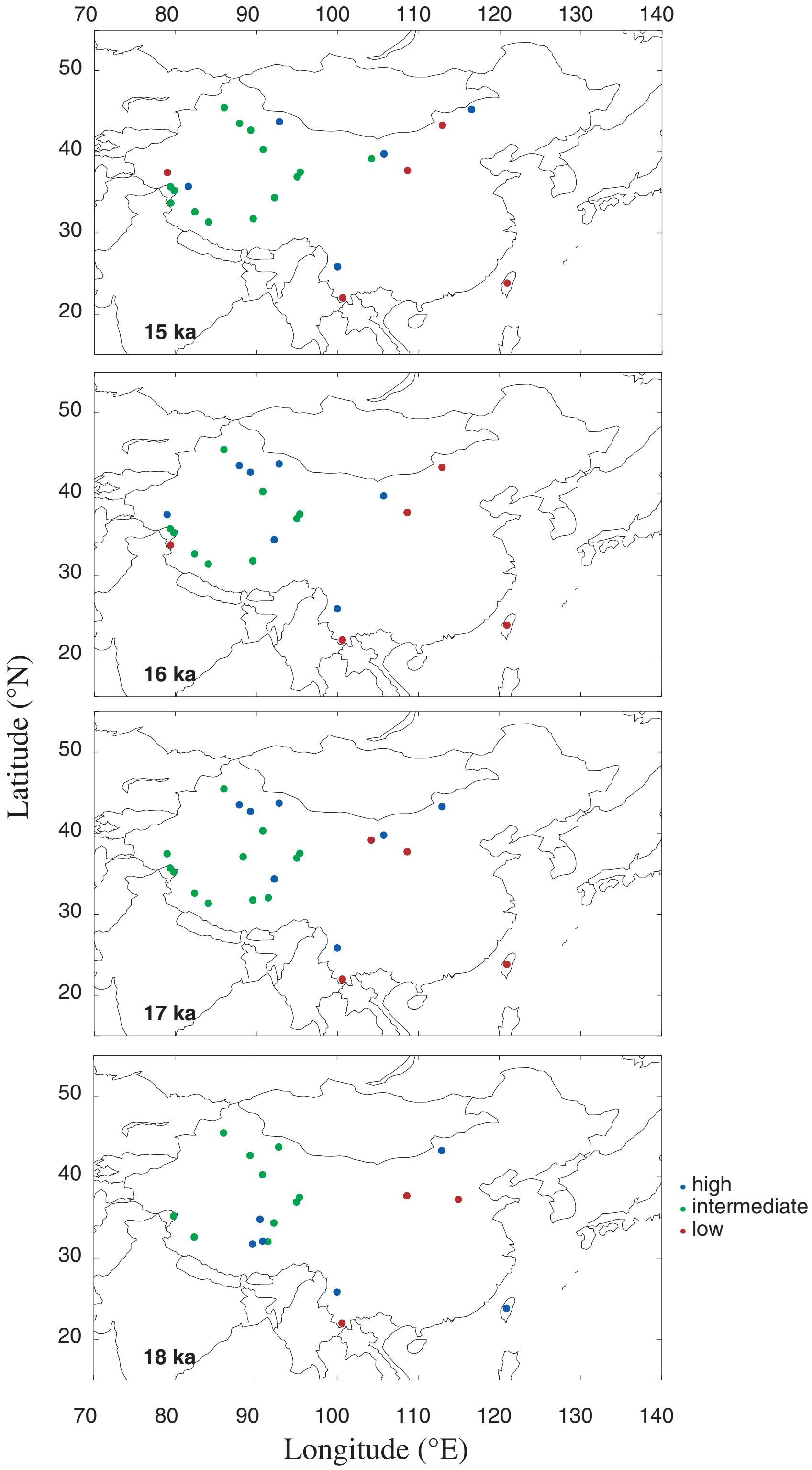
20
5 ka
6 ka
7 ka
8 ka
9 ka
Latitude (°N)
Longitude (°E)
high
intermediate
low


10 ka
11 ka
12 ka
13 ka
14 ka
Latitude (°N)
Longitude (°E)
high
intermediate
low
70
80
90
100
110
120
130
140
50
40
30
20



70 80 90 100 110 120 130 140
50
40
30
20
15 ka
16 ka
17 ka
18 ka
Latitude (°N)
Longitude (°E)
high
intermediate
low