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Palaeo-moisture evolution in monsoonal Central Asia during the last 50,000 years

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Abstract

The late-Quaternary climate history of monsoonal Central Asia was inferred from 75 palaeoclimatic records which provide information about moisture conditions in the last 50 ka (or part of this period).

Wet conditions occurred during middle and late Marine Isotope Stage 3, while the Last Glacial Maximum (LGM) was characterized by dry climate conditions in the region. A stepwise climate amelioration is suggested by the climate records following the LGM. Several climate signals of this period, which were reported from high-latitude ice core records, are preserved in archives from monsoonal Central Asia as well.

During the early Holocene, high effective moisture was inferred from most records from the area dominated by the Indian Monsoon (e.g. the Tibetan Plateau) suggesting that Holocene optimal climate conditions occurred there during this period. In contrast, areas which are dominated by the South-East Asian monsoon (SE Monsoon) and the Westerlies (in north-western and north-central China, Mongolia) do not uniformly show an early Holocene climate optimum. For this area optimal conditions prevailed during the mid-Holocene. These apparent contradictions can possibly be explained by the regional uplift and descent of air masses in the Holocene. During the early Holocene, strengthened insolation possibly caused an enhanced low-level convergence over the Tibetan Plateau which led to the intensification of the summer monsoon. The strong air uplift caused intensified precipitation and air divergence in the upper troposphere over the Tibetan Plateau. The areas adjacent to the north therefore experienced an intensified descent of air masses and consequently increased aridity. The majority of the palaeoclimatic records suggest reduced effective moisture since the late Holocene in the region.

1. Introduction

Climate models and instrumental climate records suggest that Northern Hemisphere warming is likely to have a strengthening effect on the Asian Monsoon (e.g. deMenocal and Rind, 1993), thereby possibly causing disastrous precipitation events and floods. In contrast, in the area north of the Tibetan Plateau an increasing desertification is observed which is mostly attributed to an inappropriate use of natural resources. Climate mechanisms that cause these spatial differences of dry and wet events, however, are poorly understood; a fact that was again emphasized by the most recent Intergovernmental Panel on Climatic Change report (IPPC, 2001). Therefore, the need for palaeoclimatic studies has been increasing over the last few decades since such information may help the assessment of predictions made by climate models.

The study area of this review—mainly the Tibetan Plateau, north-western and north-central China, and Mongolia—is of interest for climate studies because of two main characteristics: First, it is situated in the triangle of the Indian Monsoon, the SE Asian Monsoon

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and the Westerlies. Second, it is a region with strong climatic differences, especially concerning effective moisture. Palaeoclimatic studies therefore should yield information both on general changes in circulation systems and on the spatially different reactions to these underlying climate mechanisms.

Until now, no review of the palaeoclimatic evolution of monsoonal Central Asia has attempted to standardize and synthesize palaeoclimatic information from different archives. Such an overview is considered to be important in order to assess new data and to reach reliable conclusions about the spatial patterns of the climate history in the region.

2. Records considered and methods used

Records which provide information about the history of moisture conditions during the last 50 ka from monsoon-influenced Central Asia and adjacent areas (for simplification the area is called monsoonal Central Asia) have been compiled (Table 1). Information on moisture or precipitation given there were evaluated according to a consistent methodology (see below) to obtain an overview of the already existing palaeoclimatic data from the region. Very few records cover the complete time interval between 50 and 0 ka years BP, however, more than 84% cover at least 9000 ^{14}C-years.

Besides two (see Table 1), all profiles were dated by the radiocarbon method. Most authors, furthermore, provided their information in uncalibrated ^{14}C-years. For these reasons and since the calibration of radiocarbon dates is still a subject of permanent change the uncalibrated radiocarbon scale was used for the correlation of the given basis information. The compiled information (the effective moisture curve), however, are presented as calendar years, applying the calibration curve CalPal2003 (Weninger et al., 2004), which is mainly based on tree-ring and U/Th-coral data (Stuiver et al., 1998). All mentioned ages in the text refer therefore to calendar years (termed as ka BP). The given information about former moisture conditions was coded on a four-part scale (codes are given in brackets): dry (0), moderate dry (1), moderate wet (2), and wet (3). Each time-slice considered between 50,000 and 14,000 ^{14}C-years BP covers 500 years, while between 14 and 0 ^{14}C-years BP each single time-slice represents 100 years.

A total of 75 records were assembled and coded. Stable isotope ice-core records were not included in the data (e.g. Dunde, NE margin of the Tibetan Plateau, Thompson et al., 1990; Guliya, N Tibetan Plateau, Thompson et al., 1997), since the δ^{18}O measurements generally reflect temperature signals rather than local moisture availability.

To elucidate the regional aspects concerning the dominant circulation systems of Holocene moisture conditions, all records from northern India, northern Yunnan, and the Tibetan Plateau (west of 100° E) are classed with the Indian Monsoon (Nos. 1–27). Accordingly, all records south of the modern limit of the summer monsoon (Gao, 1962, quoted from An, 2000) and east of 100° E are classed with the SE Asian Monsoon (Nos. 28–47), which mainly comprises the Loess Plateau, the Ordos Plateau and the central Inner Mongolia province. The areas adjacent to the north (mainly Xinjiang, western Inner Mongolia and Mongolia) are generally regarded as being influenced by both the summer monsoon and—to a greater extent—by the Westerlies. Therefore, records from these regions represent the Westerlies area in Central Asia (Nos. 48–75). It is emphasized here that these artificially drawn borders represent transition zones between the circulation systems (Fig. 1).

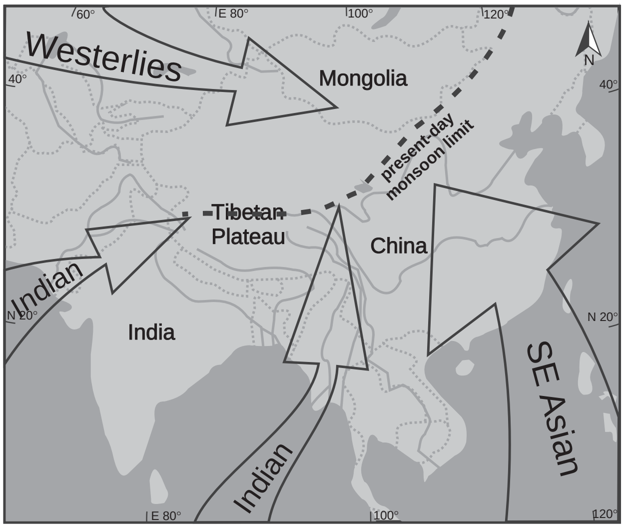


Fig. 1. Circulation systems influencing the study area (Indian Monsoon, SE Asian Monsoon, Westerlies) and present-day limit of the summer monsoon (after Gao, 1962, cit in An, 2000).

3. Results

3.1. General remarks

The locations of all records which have been included in this review are presented in Table 1 and Fig. 2. Basic information about the records (kind of archive, time coverage, sampling resolution, reliability of the chronology, proxies studied, and source) are summarized there as well. The calculated mean values of effective moisture for the whole period are shown in Fig. 3.

It should be mentioned that the general values for Marine Isotope Stage (MIS) 3 and for the Holocene cannot be exactly compared with each other, since only a few records (5) cover the whole period between 50 ka and present.

Table 1
Palaeoclimatic records from monsoonal Asia arranged according to the dominant wind system and from S to N (abbreviations see below)

Wind sys.	No.	Section	N (°)	E (°)	Elev. (m a.s.l.)	Archive	Time (cal. ka BP)	Resol. (a)	Dating No.	Dating meth.	Dating A	Dating B	Dating C	Dating D	Dating E	Methods	Reference
○	1	Qilu	24.17	102.75	1797	Lake	50.0–0	180	14	^{14}C	3	3	4	2	1	C dO G MS O	Hodell et al., 1999
○	2	Xingyun Hu	24.33	102.78	1723	Lake	25.1–0	86	20	^{14}C	2	2	4	2	1	C dO G MS O	Hodell et al., 1999
○	3	Didwana Salt L.	27.33	74.58	n.d.	Lake	23.5–0	~350	7	^{14}C	2	2	4	4	3	P	Singh et al., 1990
○	4	Lake Shayema	28.50	102.03	2400	Lake	12.9–0	100	5	^{14}C	2	2	4	3	2	P	Jarvis, 1993
○	5	Lunkaransar	28.50	73.75	n.d.	Lake	11.5–0	70	15	^{14}C	1	2	4	3	2	C dC	Enzel et al., 1999
○	6	Lake Hidden	29.82	92.38	n.d.	Lake	17.2–0	100	3	^{14}C	3	4	4	2	3	P	Tang et al., 1999
○	7	Ren Co	30.73	96.68	3120	Lake	22.3–0	260	6	^{14}C	2	3	4	2	3	P	Tang et al., 1999
○	8	Gujjar Hut	~30.9	~78.8	~4000	Peat	7.8–0	390	3	^{14}C	2	2	1	4	2	P	Phadtare, 2000
○	9	Zabuye	31.58	84.12	4421	Lake	41.1–4.5	570	3	^{14}C	4	4	4	2	3	P	Wu and Xiao, 1996
○	10	Zabuye	31.58	84.12	4421	Lake	34.8–0	740	17	^{14}C, Pb	2	2	4	2	1	C dC dO O	Wang et al., 2002
○	11	Seling Co	31.88	89.05	4530	Lake	13.8–0	220	5	^{14}C	2	2	4	2	3	P	Sun et al., 1993
○	12	Seling Co	31.88	89.05	4530	Lake	17.7–0	120	7	^{14}C	2	2	4	2	2	C G dO dC	Kashiwaya et al., 1995
○	13	No.2	~32.5	~103.3	3492	Peat	13.5–0	190	9	^{14}C	1	2	1	3	3	P	Yan et al., 1999
○	14	Hongyuan	32.77	102.50	3466	Peat	12.1–0	30	15	^{14}C	1	1	1	3	3	dC	Hong et al., 2003
○	15	Wasong	~32.9	~103.6	3490	Peat	34.0–4.5	510	9	^{14}C	2	2	4	3	3	P	Yan et al. 1999
○	16	Hongyuan	~33	~102	3490	Peat	11.5–0	430	4	^{14}C	2	2	1	4	3	P	Frenzel, 1994
○	17	Waqie	33.08	102.75	n.d.	Lake	34.0–0	—	13	^{14}C	1	3	4	3	3	S	Zhou et al., 2001
○	18	Nianbao P3	33.35	101.03	4170	Peat	10.9–0	n.d.	3	^{14}C	2	2	1	4	1	P	Schlütz, 1999
○	19	Lake Bangong	33.67	79.00	4241	Lake	11.3–0	80	25	^{14}C	1	1	3	1	1	C dC dO X	Fontes et al., 1996
○	20	Lake Bangong	33.67	79.00	4241	Lake	11.3–0	100	25	^{14}C	1	1	3	1	1	P	Van Campo et al., 1996
○	21	RM	33.95	102.35	n.d.	Peat	25.5–0	410	4	^{14}C	3	2	4	4	3	P	Shen et al., 1996
○	22	Sumxi Co	34.50	80.38	3120	Lake	14.9–0	100	6	^{14}C	2	2	4	2	3	D P Os	Van Campo and Gasse, 1993
○	23	Aksayqin Lake	35.20	79.83	n.d.	Lake	34.0–0	n.d.	21	^{14}C	1	2	4	4	4	C O P	Fang, 1991
○	24	Core 14B	36.90	100.18	3194	Lake	17.5–0	40	3	^{14}C	3	4	4	1	2	C dO	Lister et al., 1991
○	25	Q14B	36.90	100.18	3194	Lake	16.7–8.9	30	4	^{14}C	1	1	4	4	2	C dO dC MS S	Yu and Kelts, 2002
○	26	Haxi	37.50	102.40	2450	Loess	11.5–0	70	5	^{14}C	2	2	1	4	3	C MS S	Wu et al., 1998
○	27	Dunde	38.10	96.40	5325	Ice	11.6–0	~100	—	Lamina.	—	—	1	1	1	P	Liu et al., 1998
△	28	Yihai Bog	~29	~107	1774	Peat	11.3–0	40	6	^{14}C	2	3	1	4	3	P	Tong et al., 2000
△	29	Dajiu Lake 2	~32	~110	1700	Lake	14.6–0	70	7	^{14}C	2	2	4	2	3	P	Liu et al., 2001
△	30	Weinan	34.20	109.52	n.d.	Loess	28.3–12.9	200	9	^{14}C, TL	1	1	1	4	3	Mo	Wu et al., 2002
△	31	Yaoxian	34.93	108.83	n.d.	Loess	12.1–0	160	2	^{14}C	3	3	1	4	3	P	Li et al., 2003
△	32	Hezuo	35.00	102.92	~3000	Loess	50–0	100	5	^{14}C, TL	4	4	1	4	2	C G MS O	Fang et al., 2003
△	33	Yuanbo	35.63	103.17	n.d.	Loess	50–3.2	~100	5	^{14}C, TL	3	3	1	4	1	MS	Chen et al., 1997
△	34	Heimugou	35.75	109.42	n.d.	Loess	50.0–0	670	6	TL	4	4	1	4	2	MS	An et al., 1991
△	35	Dadiwan Qin'an	35.76	105.82	n.d.	Loess	12.9–0	170	4	^{14}C	2	2	1	4	3	C MS P	Xia et al., 1998
△	36	Shajinping	36.00	103.83	n.d.	Loess	50.0–0	90	11	^{14}C, IRSL	3	3	1	3	1	C G MS	Fang et al., 1999
△	37	Jingbian	37.50	102.40	3120	Loess	11.3–0	100	11	^{14}C	1	1	1	3	1	G O	Xiao et al., 2002
△	38	Midiwan	37.65	108.62	1400	Peat	15.0–10.5	60	10	^{14}C	1	1	4	3	2	dC	Zhou et al., 1999
△	39	Midiwan	37.65	108.62	1400	Peat	15.4–0	270	17	^{14}C	1	2	1	3	2	O	Zhou et al., 1996
△	40	Yangtaomao	38.80	110.45	1400	Loess	13.1–7.8	100	6	^{14}C	1	1	1	3	2	G M	Zhou et al., 1996
△	41	Lake Yanhaizi	40.10	108.45	1180	Lake	18.5–0	100	10	^{14}C	1	2	3	2	1	C E MS O	Chen et al., 2003
△	42	Dahai Lake	~40.6	~112.6	n.d.	Lake	5.8–0	90	8	^{14}C	1	1	2	2	3	C Os dO	Shen et al., 2002
△	43	Chasuqi	40.66	111.13	n.d.	Lake	10.3–0	130	4	^{14}C	3	4	4	4	3	P	Wang et al., 1997
△	44	Diaojiao Lake	41.30	112.35	1800	Lake	11.6–0	90	4	^{14}C	2	2	4	4	3	P	Shi and Song, 2003
△	45	Xiaoniuchang	42.62	116.82	1460	Lake	11.2–0	280	3	^{14}C	3	2	4	2	2	P	Liu et al., 2002a, b
△	46	Liuzhouwan	42.72	116.68	1365	Loess	16.9–0	300	2	^{14}C	3	3	4	2	1	C C/N G MS O P	Wang et al., 2001a, b
△	47	Haoluku	42.97	116.77	1295	Loess	12.1–0	190	3	^{14}C	3	3	4	2	1	C C/N G MS O P	Wang et al., 2001a, b
□	48	Hongshui River I	38.18	102.77	1460	River	31.9–10.6	160	9	^{14}C	1	1	4	4	3	P	Ma et al., 2003

Table 1 (*continued*)

Wind sys.	No.	Section	N (°)	E (°)	Elev. (m a.s.l.)	Archive	Time (cal. ka BP)	Resol. (a)	Dating No.	Dating meth.	Dating A	Dating B	Dating C	Dating D	Dating E	Methods	Reference
☐	49	Hongshui River II	38.18	102.77	1460	River	8.4–3.2	50	8	^{14}C	1	1	4	3	2	P	Zhang et al., 2000
☐	50	Sanjiaocheng	~38.	~102	1325	Lake	18.5–0	50	11	^{14}C	1	2	4	4	2	C G	Shi et al., 2002
☐	51	Yema	~38	~102	1306	Lake	15.4–0	40	4	^{14}C	2	3	4	4	2	C G	Shi et al., 2002
☐	52	Qingtu Lake	39.05	103.67	1309	Lake	6.8–0	160	6	^{14}C	1	1	4	4	3	G S	Wang et al., 1999a, b
☐	53	Baijian Lake	39.15	104.16	1280	Lake	42.4–0	—	13	^{14}C	2	2	4	4	1	S	Pachur et al., 1995
☐	54	Duantouliang	39.57	103.95	n.d.	Lake	43.7–18.5	280	9	^{14}C	1	1	4	4	2	E P Os S	Zhang et al., 2002
☐	55	W E Juyanze	41.83	101.63	898	Lake	5.5–2.7	30	4	^{14}C	1	1	4	3	1	dC dO Os S	Mischke et al., 2002
☐	56	Eastern Juyanze	41.89	101.85	892	Lake	10.6–1.5	120	5	^{14}C	1	2	4	3	1	P	Herzschuh et al., 2004
☐	57	Boston Lake	41.90	86.72	1046	Lake	8.4–0	20	5	^{14}C	1	2	1	2	2	C E O	Wünnemann et al., 2003
☐	58	Middle Tarim G2	n.d.	n.d.	n.d.	River	13.8–0	1200	5	^{14}C	2	2	4	4	3	C E S	Feng et al., 1999
☐	59	Lake Manas	45.75	86.00	251	Lake	13.8–0	~200	8	^{14}C	1	2	2	4	1	C dC dO O P X	Rhodes et al., 1996
☐	60	Lake Daba Nur	48.20	98.80	2465	Lake	13.3–0	460	6	^{14}C	2	2	4	2	3	P	Gunin et al. (eds.), 1999
☐	61	Hoton Nur	48.67	88.30	2083	Lake	11.5–0	560	6	^{14}C	2	2	2	2	1	P	Tarasov et al., 2000
☐	62	Lake Telmen	48.83	97.33	1789	Lake	7.0–0	160	6	^{14}C	1	1	3	1	1	P	Fowell et al., 2003
☐	63	Lake Telmen	48.83	97.33	1789	Lake	7.1–0	~50	6	^{14}C	1	1	3	1	1	C O S	Peck et al., 2002
☐	64	Zhalainor	49.33	117.58	n.d.	Lake	23.5–3.2	160	7	^{14}C	2	2	4	4	3	P	Yang et al., 1997
☐	65	Lake Achit Nur	49.50	90.60	1435	Lake	14.6–0	500	4	^{14}C	2	4	4	4	3	P	Gunin et al. (eds.), 1999
☐	66	Bayan Nuur	50.00	94.02	932	Lake	15.4–0	150	4	^{14}C	2	2	4	4	1	P	Grunert et al., 2000
☐	67	Bayan Nuur	50.00	94.02	932	Lake	15.4–0	—	4	^{14}C	2	2	4	4	1	S	Grunert et al., 2000
☐	68	Gun Nuur Lake	50.25	106.60	600	Lake	11.5–0	290	7	^{14}C	1	2	4	2	3	P	Dorofeyuk and Tarasov, 1998
☐	69	Gun Nuur Lake	50.25	106.60	600	Lake	11.5–0	450	7	^{14}C	1	2	4	2	3	D S	Dorofeyuk and Tarasov, 1998
☐	70	Uvs Nuur	50.37	92.20	759	Lake	16.7–0	—	3	^{14}C	3	3	4	2	3	S	Grunert et al., 2000
☐	71	Ozerki Swamp	50.42	80.47	n.d.	Peat	15.4–0	280	9	^{14}C	1	2	4	2	3	P	Kremenetski et al., 1997
☐	72	Hovsgol Lake	50.53	101.16	1645	Lake	7.0–0	470	2	^{14}C	2	2	4	2	3	P	Dorofeyuk and Tarasov, 1998
☐	73	Hovsgol Lake	50.53	101.16	1645	Lake	7.2–0	450	2	^{14}C	2	2	4	2	3	D S	Dorofeyuk and Tarasov, 1998
☐	74	Dood Nuur Lake	51.33	99.38	1538	Lake	14.6–0	660	2	^{14}C	4	4	4	2	3	P	Dorofeyuk and Tarasov, 1998
☐	75	Dood Nuur Lake	51.33	99.38	1538	Lake	14.6–0	690	2	^{14}C	4	4	4	2	3	D S	Dorofeyuk and Tarasov, 1998

Wind systems: ○—Indian Monsoon; △—SE Asian Monsoon; ☐—Westerlies

Dating reliability A: Average frequence of dating (for Holocene: 1—every 1500 a, 2—every 3000 a, 3—every 5000 a, 4—less offen; for Pleistocene:—every 5000 a, 2—every 10,000 a, 3—every 20,000 a, 4—less offen).

Dating reliability B: Maximum interval of datings (for Holocene: 1—2000 a, 2–4000 a, 3–6000 a, 4—more; for Pleistocene:—10,000 a, 2–20,000 a, 3–30,000 a, 4—more).

Dating reliability C: Old carbon reservoir effect (1—no influence possible, 2—reservoir effect checked (but less than 500 a), 3—reservoir effect checked, 4—reservoir effect not checked).

Dating reliability D: Continuality of the record (1—no or only slight changes in the sediments, no interruptions, 2—sediment changes, no interruptions, 3—interruptions possible, covered by datings, 4—no attempt made to interruptions).

Dating reliability E: Presentation of the time model in the publication (1—presentation of each single dating result with additional information e.g. lab.-No., δ^{13}C-value, extesiv discussion of the time model, 2—presentation of the original dating results, discussion of the time model, 3—presentation of the original dating result, 4—no presentation of the original dating result).

Methods: Underlined methods are considered by the time resolution; C—carbonate content; C/N—carbon/nitrogen ration; D—diatoms; dC—carbon stable isotope; dO—oxygen stable isotope, E—element conc.; G—grain-size; M—minerals; Mo—molluscs; MS—magnetic susceptibility, O—organic content; Os—ostracods, P—pollen; S—sediment description; X—X-ray diff.; lam.—laminations).

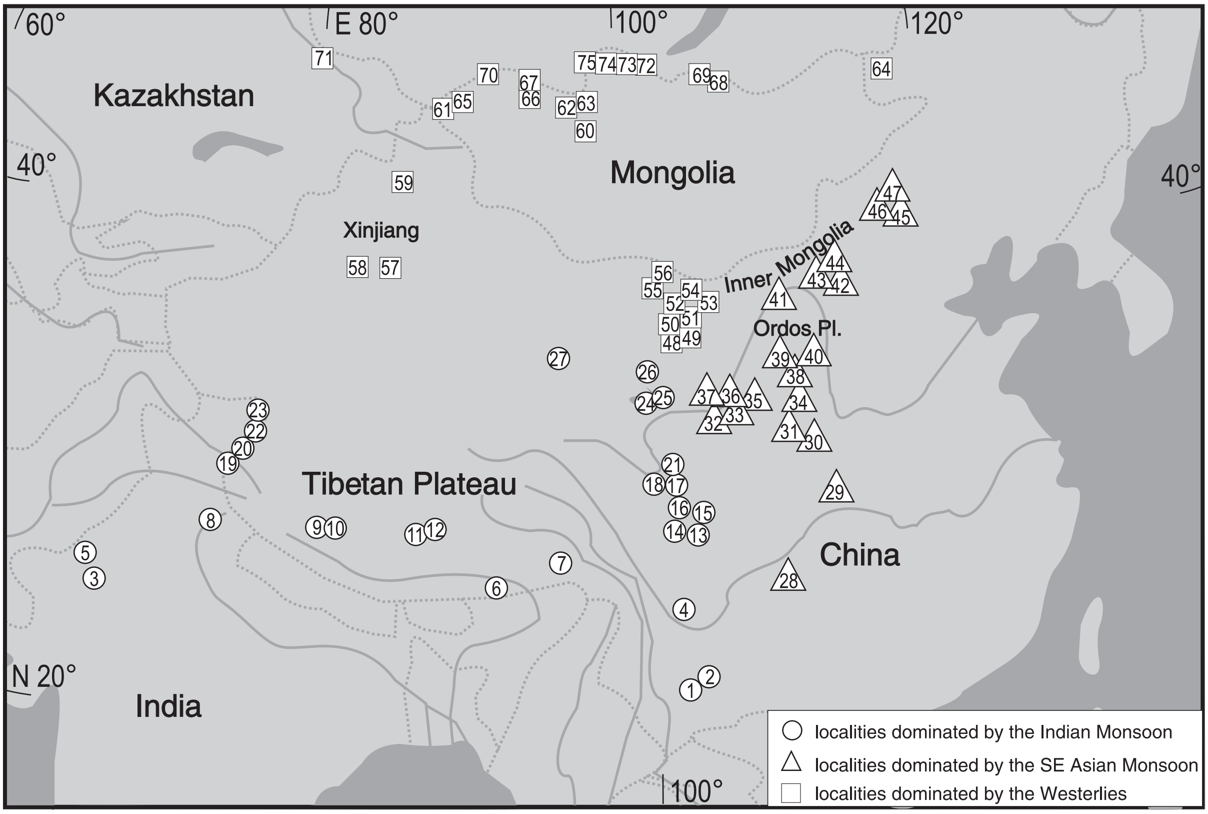


Fig. 2. Spatial distribution of palaeoclimatic studies (numbers indicated refer to the records, see Table 1) from monsoonal Central Asia that are included in the synthesis; records are numbered according to the dominant wind system and from S to N.

3.2. MIS3 und MI2

The middle and late MIS 3 yielded values between moderate dry and wet (Fig. 4). However, only a few well-dated records are currently available. Most favourable moisture conditions are suggested between 43 and 37.6 ka BP when more than 70% of the records indicate moderate wet or wet conditions. After 25.5 ka BP the mean moisture values decrease noticeably (for >70% moderate dry or dry conditions are recorded), indicating reduced moisture in the region during MIS 2. The moisture minimum of the entire record was found between 21.3 and 19.8 ka BP, when uniformly all records suggest dry (74%) or moderate dry (26%) conditions. Between 19.8 and 17.2 ka BP the mean moisture values increase gradually. More than 40% still indicate dry conditions, however. Stable or even slightly reduced moisture is suggested between 17.2 and 15.4 ka BP. A sharp increase in effective moisture is displayed in the subsequent period (15.4–13.0 ka BP), as most records now show moderate dry ~50% or moderate wet (~30%) conditions. The period between 13.0 and 11.6 ka BP is marked by a return towards lower moisture availability again. Moisture conditions became significantly wetter at the Pleistocene/Holocene transition at 11.5 ka BP.

3.3. Holocene

The mean values of all records show optimal moisture conditions during the first half of the Holocene, while later on, especially after 3.2 ka BP, markedly drier conditions are recorded in monsoonal Central Asia. An assignment of the records to the dominating circulation system—Indian Monsoon, SE Asian Monsoon, Westerlies—provides a more detailed insight into the moisture evolution. The calculation of the mean effective moisture of each region and the percentage frequencies, as displayed in Fig. 5, yield distinct differences between the areas of the single circulation systems, especially concerning the period of the Holocene optimum (when >50% of all records of the area indicate wet conditions).

The Indian Monsoon area indicates optimal moisture conditions during the early Holocene (10.9–7.0 ka BP) and shows rather wet conditions (when >50% of all records yielded moderate wet or wet) until 4.3 ka BP. The SE Asian Monsoon area shows rather wet conditions from early until the middle of the late Holocene (11.5–1.7 ka BP), but in contrast to the Indian Monsoon area it shows optimal moisture during the early mid-Holocene (8.3–5.5 ka BP). The Westerlies area does not display a pronounced moisture maximum (always <50% of sites indicated wet conditions) but rather showing almost constant values between 12.1 and 2.7 ka BP, with a small maximum

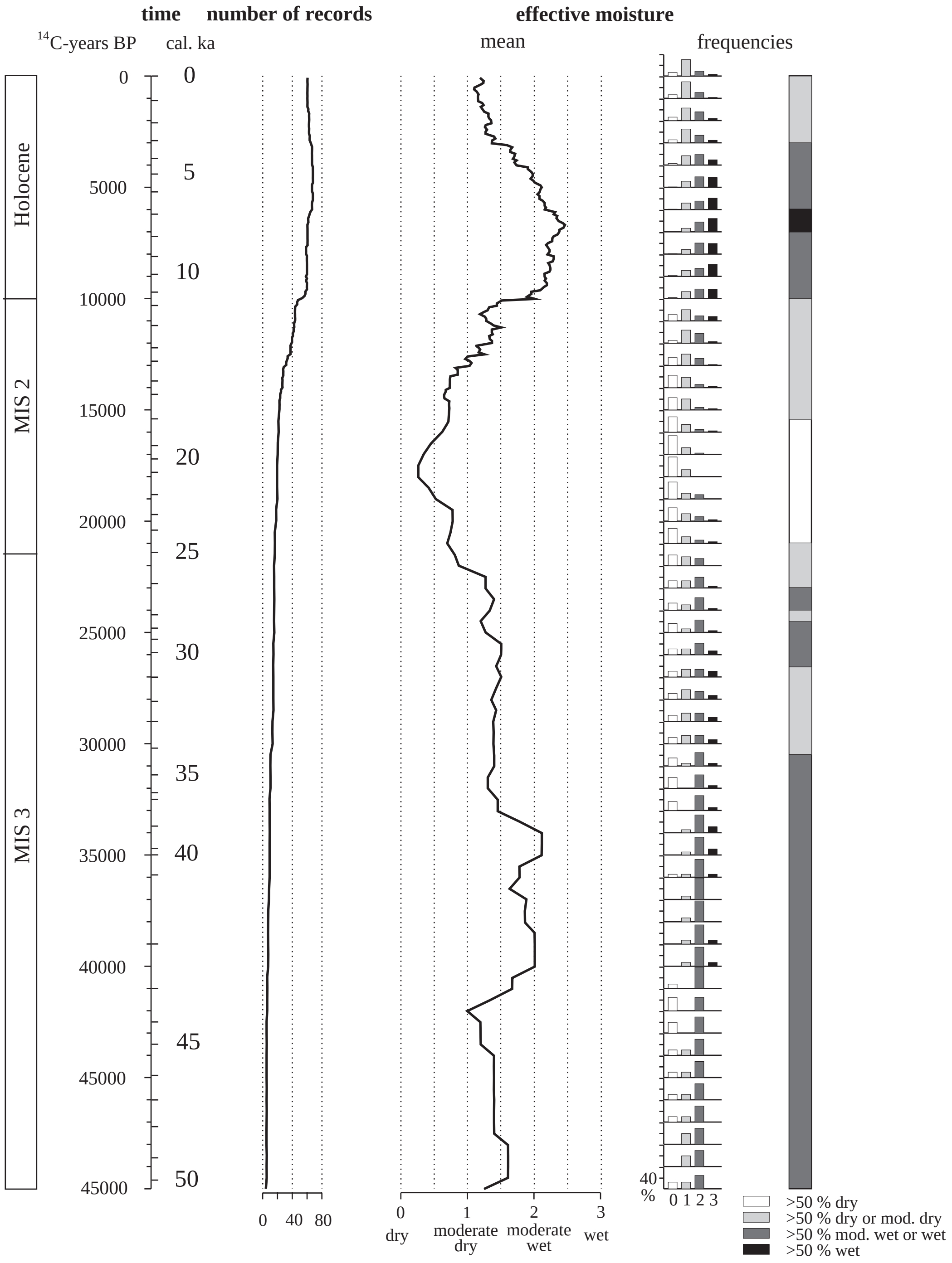


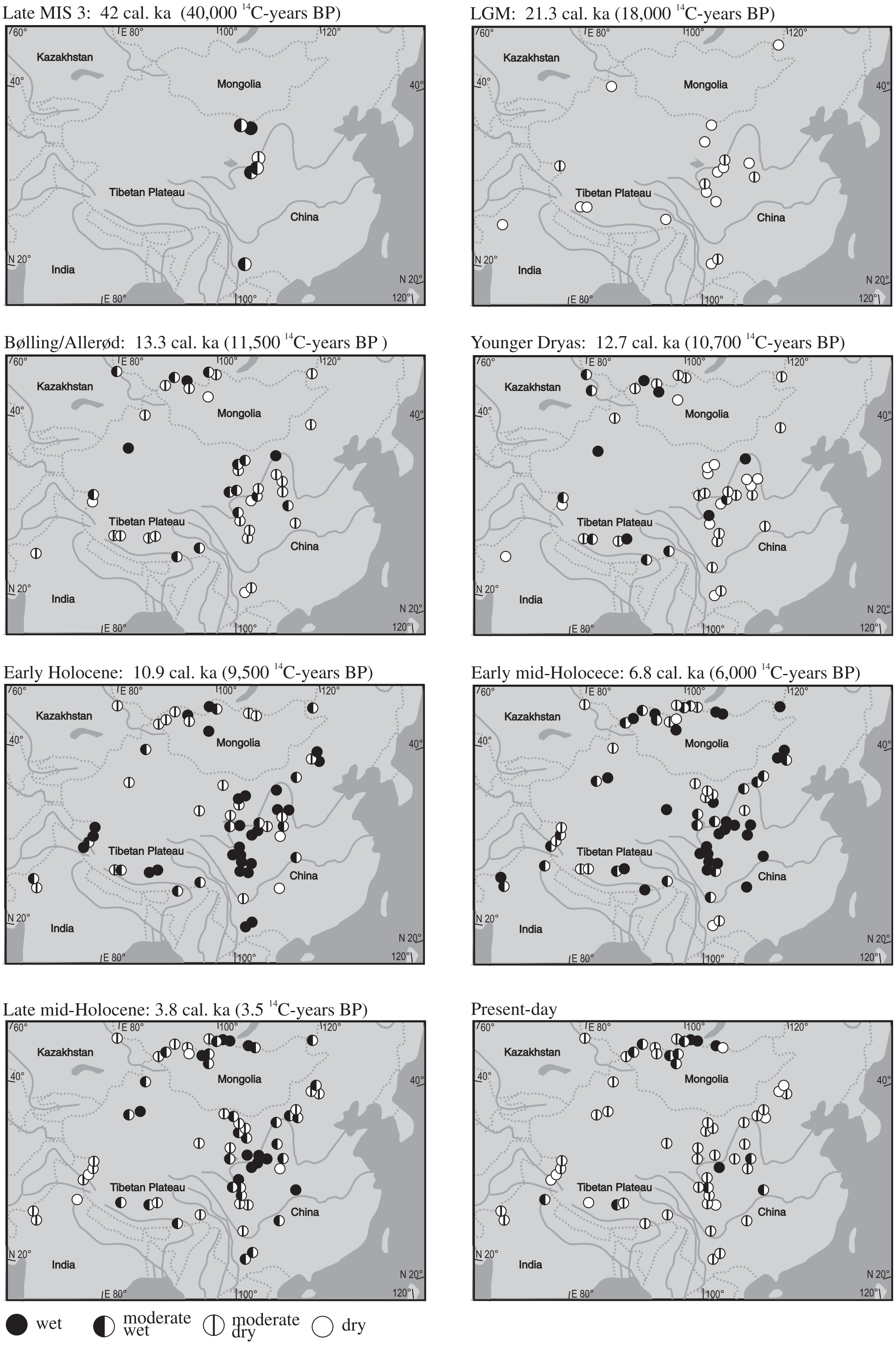
Fig. 3. Mean effective moisture and percentage frequencies of recorded moisture from monsoonal Central Asia. The left curve indicates the number of records available for each single time slice.

between 7.5 and 6.8 ka BP, when more than 40% of the records indicate wet conditions. The early Holocene yielded inconsistent results of simultaneously rather dry, rather wet, and wet conditions in the area. In contrast to the monsoon-dominated areas, however, records with rather dry conditions predominate in the Westerlies area, especially during the first two thousand years of the Holocene. The curves of all three areas uniformly display that mean effective moisture decreased after ~3 ka BP.

4. Discussion and conclusions on the moisture evolution

4.1. *Forcing mechanisms of monsoon activity between 50 and 0 ka BP*

For more than a decade, the scientific community has been extensively discussing whether changes in the insolation or changes in the glacial boundary conditions (ice volume, sea surface temperature (SST), albedo)



Late MIS 3: 42 cal. ka (40,000 ^{14}C-years BP)

LGM: 21.3 cal. ka (18,000 ^{14}C-years BP)

Bølling/Allerød: 13.3 cal. ka (11,500 ^{14}C-years BP)

Younger Dryas: 12.7 cal. ka (10,700 ^{14}C-years BP)

Early Holocene: 10.9 cal. ka (9,500 ^{14}C-years BP)

Early mid-Holocece: 6.8 cal. ka (6,000 ^{14}C-years BP)

Late mid-Holocene: 3.8 cal. ka (3.5 ^{14}C-years BP)

Present-day

Fig. 4. Spatial patterns of effective moisture from monsoonal Central Asia for single time slices.

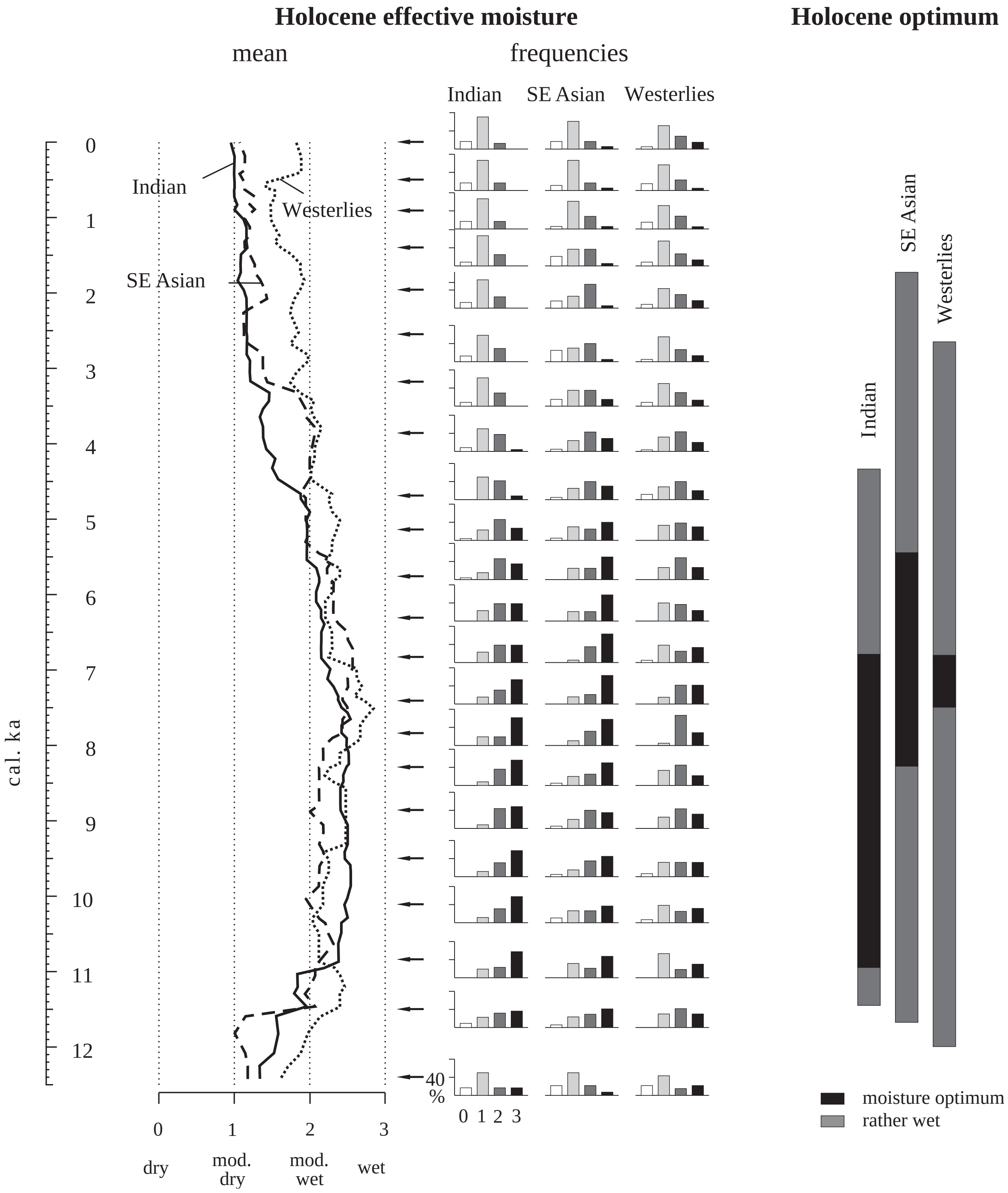


Fig. 5. Classification of the palaeoclimatic records from three regions, that represent single circulation systems; the curves indicate regional differences in the Holocene effective moisture (mean, frequencies). Holocene moisture optimum is assumed to occur when more than 50% indicate wet conditions in the Indian and SE Monsoon areas; and more than 40% indicate wet conditions in the Westerlies area.

influenced the millennial-scale monsoon activity to a significant extent (e.g. Clemens and Prell, 1991; Overpeck et al., 1996; Fleitmann et al., 2003).

On a rough timescale, the late-Quaternary monsoon history derived from the compiled records has a similar evolutions like the insolation changes for 30° N in June (Berger and Loutre, 1991; Fig. 6) showing strong insolation and high effective moisture at the end of the MIS 3 and during the early Holocene in monsoonal Central Asia, while during the periods of weak insolation the study area experienced rather dry conditions (around the Last Glacial Maximum (LGM) and during the Late Holocene). However, the exact peaks do not exactly coincide. Whether these differences is caused by other monsoon driving factors or is only pretended by chronological inconsistencies (which could be the case especially for the MIS 3) could not be clarified here.

Palaeoclimate studies on sediments from the Arabian Sea recently revealed also that the Indian Monsoon is highly sensitive to insolation differences on millennial, centennial and decadal time scales (Agnihotri et al., 2002; Leuschner and Sirocko, 2003).

Many authors, however, have stressed the central role of glacial boundary conditions in modifying the response of the monsoon to astronomical forcing (e.g. Overpeck et al., 1996; Sirocko et al., 1993; Fleitmann et al., 2003). Accordingly, the $\delta^{18}O$ curve of the Guliya ice-core from the west Kunlun Mountains on the northern margin of the Tibetan Plateau, indicates that temperature fluctuations during the last 125 ka are consistent with glacial cycles recorded in Northern Hemisphere ice-cores and North Atlantic Ocean sediments (Thompson et al., 1997; Yao et al., 1997, 2000). Ocean sediment and loess archives have also provided similar evidence (e.g. Porter and An, 1995; Chen et al., 1997; Schulz et al., 1998; Wang et al., 1999a, b; Porter, 2001; Fang et al., 2003). Accordingly, the general climate changes during the last 50 ka which have been deduced from the GRIP ice-core record (Johnsen et al., 2001) are also reflected in the mean effective moisture curve from the monsoonal Central Asia (Fig. 6). Besides long-term climate cycles, even millennial scale climate events like the Younger Dryas and the Heinrich 1 event appear to be recorded for the Holocene and the MIS 2 section. While, such short-term events are not traced during the MIS 3 possibly due to poor dating control and low data resolution of the records.

4.2. *Moisture evolution during late MIS 3 and MIS 2*

Only a few climate records (mostly from the Loess Plateau and the Tengger Desert) are available from monsoonal Central Asia for this period, most of them suggesting moderate wet conditions (Fig. 4; centred around 40 ka BP). More information on wet climate conditions for 40–30 ka BP, especially for the Tibetan Plateau, have been deduced either from dated exposed lake sediments or from analysed borehole samples, or have been derived from profiles with poor chronology and low sample resolution. Summaries of these potentially useful information are provided by Li et al. (1991), Shi et al. (1999, 2001), Li (2000), and Shi and Yu (2003).

The precessional cycle for 30° N reveals strong summer insolation during ~40–30 ka BP, which exceeds the present-day conditions in the area and nearly equals early Holocene values (Berger and Loutre, 1991). Accordingly, several studies suggest that the summer monsoon was strong during the late MIS 3, but except for the Tibetan Plateau not as strong as it was during optimal Holocene periods (Fang et al., 1999, 2003; Hodell et al., 1999; Wang et al., 1999a, b). Due to the persistent large Northern Hemisphere ice-volume the glacial boundary conditions might have strongly influenced the monsoon system during MIS 3 (deMenocal and Rind, 1993). Therefore, monsoon activity possibly could not respond to insolation in a similar way as it did during the early Holocene (Hodell et al., 1999). Several results from the Tibetan Plateau, however, are at variance with a less active summer monsoon during late MIS 3. Shi et al. (2001) rather proposed temperatures 2–4 °C higher and precipitation 40–100% higher than today based on the $\delta^{18}O$ values of the Guliya ice-core record (Thompson et al., 1990, 1997). Further evidence of a comparatively wet period on the Tibetan Plateau at this time comes from dated lake sediments, which indicate lake levels above the present-day conditions for example, in western Tibet at Tianshuihai (Li et al., 1991), Akesaiqin (Fang, 1991), Bangong Co (Li et al., 1991) and in central Tibet at Zabuye Lake (Wang et al., 2002) and Selin Co (Li et al., 1991). Large lakes, which have been reported for the Qaidam Basin (Chen and Bowler, 1986) and the Alashan Plateau (Pachur et al., 1995; Wünnemann et al., 1998a, b; Zhang et al., 2002) do not prove that similar wet conditions prevailed also in other areas of north-west China, since these lakes have been fed by rivers which originate in the Tibetan Plateau or in bordering mountain ranges. The occurrence of such extended lakes, however, is further evidence for strongly increased precipitation values over the Tibetan Plateau area itself.

Generally, a strongly intensified winter monsoon and a weakened summer monsoon is assumed for the climate of MIS 2 (~25–11.5 ka BP, e.g. An et al., 1991; Chen et al., 1997; Wang et al., 1999a ,b). Most records from the area yield dry or moderate dry conditions for the LGM at ~21 ka BP. Some workers have, however, proposed higher effective moisture conditions during MIS 2 in western China because of less strong evaporation at lower temperatures and displacements of the circulation systems (Shi et al., 1997; Qin and Yu, 1998; Liu et al., 2002a, b; Shi, 2002; Yu et al., 2003). During the period of the Last Termination the mean effective moisture from monsoonal Central Asia (Fig. 6) matches the $\delta^{18}O$ curve of the GRIP ice core from Greenland (Johnsen et al., 2001). Subsequent to the insolation minimum at ~23 ka BP, the insolation steadily increased during the Last Termination and reached its maximum at the Pleistocene–Holocene transition (Berger and Loutre, 1991). In accordance with the GRIP record (Johnson et al., 2001) many sequences from the investigated area display climate amelioration following the LGM. A

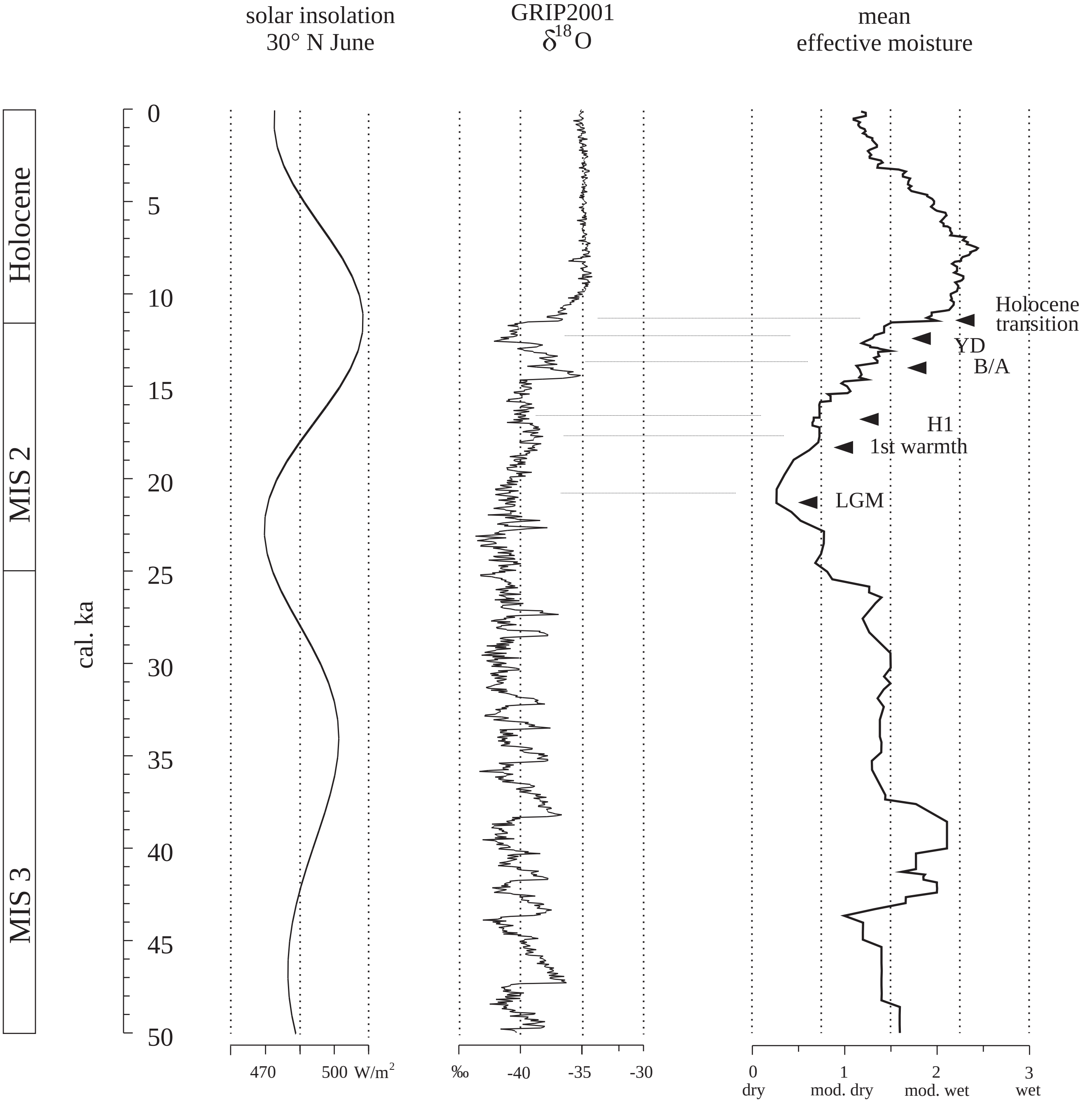


Fig. 6. The mean effective moisture from the Asian monsoon margin in comparison to the insolation at 30° N (source: Berger and Loutre, 1991), and to the GRIP δ^{18}O record from Greenland (source: Johnsen et al., 2001).

phase of stable and slightly wetter conditions in Asia is indicated between 18.5 and 17 ka BP. It possibly represents the onset of summer monsoon circulation after the LGM. It is hard to identify, however, whether this is a result of increased monsoonal precipitation reaching the interior of Asia or if it is a result of local glacier melting. Slightly drier conditions are found in some records (e.g. Chen et al., 1997) between 17.2 and 15.4 ka BP, which is displayed as a cooling phase in the GRIP record and matches the Heinrich 1 event in the North Atlantic (Bond et al., 1992). Many records from monsoonal Central Asia demonstrate that the first strong intensification of the Asian monsoon circulation after the LGM, took place at the beginning of the Bølling/Allerød period (e.g. Zhou et al., 1999; Wang et al., 2001a, b). Overpeck et al. (1996) consider the abrupt warming of the North Atlantic during this period to be the reason for this (see citations ibid.). According to their interpretations, this high-latitude warmth led to a slightly greater warming over the Tibetan Plateau and a reduction of the snow cover, which in turn gave rise to a stronger Indian Monsoon. The Bølling/Allerød warm phase between 14.8 and 12.8 ka BP shown in the GRIP record is mirrored in the Asian records as well, indicating that the period after 13.8 ka BP (~Allerød) was slightly wetter. The Younger Dryas event (~12.8–11.5 ka BP), reflected as a cold event in Greenland, yields drier climate conditions in areas from monsoonal Central Asia (Fig. 6), especially on the Loess Plateau (e.g. Chen et al., 1997).

4.3. Moisture evolution during the Holocene

Continental records from across monsoon-influenced Asia (e.g. Zhou et al., 1996; Enzel et al., 1999; Fleitmann et al., 2003; Hong et al., 2003; Yuan et al., 2004) and studies from the bordering oceans (e.g. Sirocko et al., 1993; Wang et al., 1999a, b) indicate that a strong intensification of both the Indian and the SE Asian Monsoon occurred at the Pleistocene–Holocene transition at ~11.5 cal. ka. Comparison of the mean effective moisture from monsoonal Central Asia to the GRIP record (Johnsen et al., 2001) reveals that the abrupt monsoon intensification was in phase with rising air temperature in the northern Atlantic region (Fig. 6).

Almost all high-resolution marine palaeoclimate archives from both the Arabian Sea (e.g. Sirocko et al., 1993; Overpeck et al., 1996) and from the South China Sea (e.g. Sun and Li, 1999; Wang et al., 1999a, b) indicate that summer monsoon in low-latitude marine areas was strongest during the first half of the Holocene (~11.5–6 ka BP). Except for stalagmite records from southern Oman (Fleitmann et al., 2003) and southern and eastern China (Wang et al., 2001a, b; Yuan et al., 2004), continental records from monsoonal Asia do not equal these marine records in time resolution and dating control. Therefore, the marine records are essential to understand the underlying mechanisms which caused climatic changes on the Asian continent. In contrast to the marine records, not all records from monsoonal Central Asia indicate optimal moisture conditions at the time immediately following the Pleistocene–Holocene transition (Fig. 4). There exist rather extreme differences in the recorded Holocene optimum even when the investigated sites are located close to each other. For example, palaeoenvironmental reconstructions derived from the Eastern Juyanze record (Hartmann et al., 2003; Herzschuh et al., 2004) suggest that dry conditions with desert vegetation and low lake levels prevailed on the Alashan Plateau during the first half of the Holocene, while the Holocene optimum occurred during the late mid-Holocene. In contrast, in the central Qilian Mountains, which are located ~400 km further to the south, the highest lake levels and maximum forest vegetation were reconstructed for the early Holocene, while the site is characterized by lake-level lowering and alpine vegetation since the mid-Holocene (Herzschuh et al., 2005; Mischke et al., 2005). Likewise, highest lake levels and favourable vegetation conditions are reported from other sites on the Tibetan Plateau for the early Holocene (e.g. Sumxi Co: 11.5–7.5 ka BP, Van Campo and Gasse, 1993; Seling Co: 10.9–7.5 ka BP, Sun et al., 1993; Zoige Basin: 10.9–5.5 ka BP, Hong et al., 2003). Palaeoclimate studies from north-central and northern China commonly indicate a mid-Holocene optimum (Shi et al., 1993; An, 2000; e.g. Dajiu Lake in Hubei: 8.4–3.6 ka BP, Liu et al., 2001; Chasuqi section in Central Inner Mongolia: 5.8–4.7 ka BP, Wang et al., 1997). Lake archives from the Thar desert in India also point to the most humid conditions during the early mid-Holocene (Lake Didwana: 8.3–7.3 ka BP, Singh et al., 1990; Lake Lukaransar: 7.1–5.8 ka BP, Enzel et al., 1999). Compilation of paleaoclimate proxy data from Mongolia suggests an early Holocene rather dry phase, while more wet conditions are suggested for the mid-Holocene period (Harrison et al., 1996; Tarasov and Harrison, 1998).

Inconsistencies in the appearance of the humid phase can partly be explained by poor sample resolution and an inadequate dating quality of the different records. Such a simple explanation, however, cannot solve the problem of asynchronous climate changes in the whole region.

According to a general palaeoclimatological assumption, the area should very sensitively reflect the spatial expansion or withdrawal of the dominating circulation systems during the Holocene because of its marginal position in relation to the summer monsoon and the Westerlies (Chen and Holmes, 2003). According to this assumption, higher moisture values on the Tibetan Plateau during the early Holocene can possibly be attributed to an enhanced Indian Monsoon, while the areas to the north and north-east were solely influenced by the dry Westerlies and a weak SE Asian summer monsoon during this period. This cannot, however, explain why the strengthened Indian Monsoon did not extend its influence further to the north. Furthermore, comparison of marine records from the South China Sea and the Arabian Sea indicate that both short-lasting events and long-term variations in the SE Asian Monsoon were coeval with variations in the Indian Monsoon regime (Wang et al., 1999a, b). In conclusion, regional differences concerning humid phases on the continent therefore are very unlikely to be due to differences between circulation mechanisms.

Another explanation for wet and warm conditions on the Tibetan Plateau during the early Holocene and for simultaneously prevailing dry conditions in the area adjacent to the north might relate to regional differences in the uplift and descent of air masses. The strong insolation in summer gives rise to strong uplift of air masses (which causes latent heat release and hence precipitation in the area) forming the a low pressure cell (termed the Tibetan Low). This leads to a large-scale convergence (cyclonic circulation) in the lower troposphere over the Tibetan Plateau and consequently to an anticyclonic circulation and divergence in the upper troposphere. This circulation brings enhanced subsidence of air masses to the lowland areas adjacent to the north of the Tibetan Plateau, which increases the aridity in the area (Broccoli and Manabe, 1992). The strength of these circulation patterns likely follows the strength of the insolation and hence the monsoon activity. As a consequence, stronger monsoon activity over the Tibetan Plateau during the early and mid-Holocene in combination with increased precipitation would cause enhanced aridity in the lowland areas further to the north. This pattern is suggested by the compiled records from the area which indicate favourable moisture conditions in the Plateau areas and predominantly arid conditions in the lowlands of Inner Mongolia, Xinjiang, and Mongolia. Furthermore, this explanation is supported by results from a numerical general circulation model (GCM) experiment performed by Bush et al. (2002). The model (global, fully coupled atmosphere-ocean GCM), which among other parameters predicts soil moisture, has been configured for 6 ka BP and for today in order to explain paleaeoclimatic patterns on the Loess Plateau. In Bush et al. (2002) the predicted spatial distribution of soil moisture is given. This parameter is considered to reliably represent effective moisture. For the mid-Holocene simulation, the model yielded increased soil moisture on the Tibetan Plateau and the Loess Plateau as a consequence of the enhanced summer monsoon, whereas the areas adjacent to the north (including Mongolia) show drier land surfaces compared to present-day conditions as a consequence of strong air mass subsidence.

Climatic changes in monsoon-influenced Asia are generally assumed to appear as a combination of cold and dry or of wet and warm conditions in response to the weakening or strengthening of the summer monsoon, respectively (e.g. Yao et al., 1996). However, different combinations have also been reported, but they were attributed to strong fluctuations and climate instability (e.g. Zhou et al., 2001; Wu et al., 2002). Most Holocene continental climate archives in Asia are lake sediments which permit the reconstruction of lake-level changes and vegetation dynamics. Climate changes are therefore reflected as changes in the effective moisture, which represents a combined signal of temperature and precipitation changes. Especially in extreme arid regions, a stronger summer monsoon with irregular precipitation should therefore result in less effective moisture due to enhanced evaporation. In contrast, in humid areas the evaporation could only slightly increase since the air is already fully saturated with water vapour. A stronger monsoon signal therefore results in rising lake levels and increasing available moisture for plant growth. It can thus be concluded, that a stronger summer monsoon during the first half of the Holocene would result in decreased effective moisture in the arid areas of Inner Asia and more favourable conditions in humid areas especially on the eastern Tibetan Plateau, which is what has been revealed by this study.

This interpretation is supported by results derived from biome reconstruction for 6.8 and 0 ka BP on the basis of pollen data (Yu et al., 1998). During the early and mid-Holocene, forests in eastern Tibet expanded into high elevations where alpine plant communities occur today. This points to an increased growing-season warmth. Furthermore, the extension of temperate deciduous forests further to the north-east, far beyond their present limits, points to higher temperatures during this period as well, since the forests depend on mild winters. In contrast, the desert areas in northern and western China do not yield discernible improvement of the growing conditions between 6.8 ka BP and present-day. It should be kept in mind, however, that the region is represented only by a few data points.

The results presented here suggest that regional differences in the Holocene climate optimum cannot be attributed to boundary shifts in the circulation systems. They are probably caused by single climate signals (e.g. an enhanced summer monsoon in the early Holocene) that gave rise to different moisture availabilities in more humid and arid areas due to the interaction between precipitation and temperature. Therefore, this synthesis calls into question the idea that the semi-arid and arid areas in northern China are suitable for the reconstruction of general monsoon mechanisms. It seems more likely that palaeoclimatic studies in these areas offer the opportunity to understand spatially varying reactions on the continent to changes in the underlying climate mechanisms, which nevertheless open an exciting and challenging prospect.

The gradual monsoon weakening since the mid-Holocene, which is suggested by marine records, is generally interpreted as a response to declining summer insolation (e.g. Overpeck et al., 1996; Gupta et al., 2003). Especially since 3 ka BP most records from the areas dominated by the Asian monsoons show markedly lower effective moisture, while in the Westerlies-dominated areas no uniform decline in the effective moisture can be deduced.

What conclusions can be drawn from this study concerning recent regional climate change? The Chinese government has paid much attention to the problem of desertification. However, the environmental changes in the deserts of northern China and the underlying mechanisms of these changes are still poorly understood. Since the last few hundred years an intensification of the Indian Monsoon activity is reported to be in phase with temperature increase in the Northern Hemisphere (Anderson et al., 2002). By analogy with the early Holocene processes discussed here, further human-induced global warming would result in a further strengthening of the Indian Monsoon which would even enhance the aridity in the lowlands of Central Asia. Therefore possibly the problem of recent desertification in Central Asia may result not only from inadequate land-use but also from an enhanced greenhouse effect.

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