QUATERNARY CLIMATE CHANGE IN IRAN – THE STATE OF KNOWLEDGE

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With 5 figures and 1 table
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Summary: The mostly Mediterranean climate of Iran is governed by the pressure systems of the Siberian High, the Westerly depressions and the SW Monsoon. In the past, the locations and intensities of these systems changed probably causing climate change and affecting landscape evolution in this ecologically diverse country. Recently, new evidence for Quaternary climate change in Iran has been presented. This paper briefly reviews the present state of knowledge and identifies future perspectives of paleoclimatic research in Iran. Paleoclimatic deductions have mainly been based on geomorphic evidence and, more recently, also on stratigraphical investigations including the physical dating of sediments. In northern and western Iran climate changed between dry and cold climatic conditions during the stadials and moist and warm conditions during the interglacials. Lake sediments and loess deposits also suggest moisture increases during interstadials of the Last and Penultimate Glacials. In western Iran, the Younger Dryas and the Lower Holocene were most probably characterized by dry climatic conditions. Overall, the climatic cycles and events known from other parts of the globe are rarely documented in Iran and our picture of past climate change there is patchy and incomplete. More proxy-information and geochronological data are needed, in particular for central and southern Iran. The sedimentary records of lakes and playas as well as loess deposits hold a strong potential to identify climate signals and the paleoclimate information of speleothems or tree-rings has not yet been challenged.

Zusammenfassung: Das im Wesentlichen mediterrane Klima des Iran wird durch die Drucksysteme des Sibirischen Hochs, der zyklonalen Depressionen und des SW Monsuns bestimmt. Die Intensitäten und Positionen dieser Drucksysteme variierten in der geologischen Vergangenheit, was zu Klimawechseln und Folgen für die Landschaftsgenese in diesem naturräumlich vielseitigem Land geführt haben wird. Unlängst wurden neue Belege für quartäre Klimawechsel in Iran präsentiert. Dieser Beitrag stellt den aktuellen Kenntnisstand zusammen und zeigt Perspektiven für künftige paläoklimatische Forschungen auf. Schlussfolgerungen über das Paläoklima Irans wurden bisher im Wesentlichen aus geomorphologischen Hinweisen gewonnen und in jüngerer Zeit auch aus stratigraphischen Untersuchungen und physikalischen Altersdatierungen von Sedimenten. Demnach wechselte das Klima in Nord- und Westiran von trocken-kalten Bedingungen während der Stadiale zu feucht-warmen Verhältnissen während der Interglaziale. Seesedimente und Lössablagerungen deuten auch Phasen vergleichsweise feuchterer Bedingungen während des letzten und vorletzten Glazials an, die wahrscheinlich mit Interstadialen zu korrelieren sind. In Westiran waren die Jüngere Dryas und das untere Holozän sehr wahrscheinlich durch trockene Klimaverhältnisse gekennzeichnet. Insgesamt sind die aus anderen Regionen bekannten Klimazyklen und -ereignisse noch wenig dokumentiert und unser Bild des Klimawandels in der Region ist daher sehr lückenhaft und unvollständig. Vor diesem Hintergrund werden mehr Proxy-Informationen und geochronologische Daten benötigt, insbesondere über Zentral- und Südiran. Die Sedimentarchive der iranischen Seen und Playas sowie Lössablagerungen bieten ein großes Potential, weitere Klimasignale zu identifizieren. Dies gilt auch für die noch kaum untersuchten Klimaarchive von Speleothemen und Baumringen.

Keywords: Iran, climate change, Quaternary, geomorphic evidence, physical dating, lake sediments, loess, pedostratigraphy

1 Introduction

The Quaternary has been a period of global cooling and pronounced climate changes from interglacial to glacial and interstadial to stadial times as, for instance, documented in Oxygen isotope values ($\delta^{18}O$) of deep sea sediments and ice cores. During the Milankovitch chron (Berger 1994), the timing of climate change has correlated with the variations of the sun-earth geometry in cycles of 100.000 yr (100 ka), 40 ka and 23 ka, related to changes in excentricity, obliquity and precession of the earth's orbit, respectively. In addition, millennial-scale oscillations in $\delta^{18}O$ values, the so-called Dansgaard-Oeschger events (Dansgaard et al. 1993; Johnsen et al. 1992) have occurred, documenting slow cooling phases at

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the beginning of a stadial and fast temperature rises at the start of an interstadial. The causes of these oscillations might be deviations in ocean surface currents, surges of ice sheets, variations in sunspot activity or instabilities in the atmospheric carbon dioxide (CO_2) system. Although the Dansgaard-Oeschger events have been recorded in many different places (Voelker 2002), it is not clear whether these short-term climatic fluctuations occurred worldwide. The spatial extent of other climate signals, such as the Heinrich events or the 8.2 ka event (Rohling and Pälike 2005), is also insufficiently known.

The frequency, direction and timing of Quaternary climate change in Iran and western Asia as a whole are sparsely documented, compared to the extensive literature on other parts of the globe. Climatic conditions in Iran are mainly governed by the pressure systems of the westerly cyclones, the Siberian High and the SW-Monsoon. During the Quaternary these systems possibly changed their relative locations and intensities, and thus their influence on climate varied. In order to reconstruct these changes, evidence of Quaternary climate change in Iran is needed. This will also help to improve our understanding of landscape evolution in Iran and in the old-world dryland belt as a whole.

Evidence of Pleistocene and Holocene climate fluctuations in Iran has been deduced from investigations of different geomorphic systems including lake sediments (e.g., Van Zeist and Wright 1963; Van Zeist and Bottema 1977; Kelts and Shahrabi 1986; Djamali et al. 2008; Ramezani et al. 2008), playas or kavirs (Krinsley 1970), glacial moraines (e.g., Bobek 1937; Hagedorn et al. 1975; Kuhle 1976, 2008; Preu 1984) and periglacial features (e.g., Hagedorn et al. 1978), salt domes (Busche et al. 2002), alluvial sediments (Scharlau 1958; Vita-Finzi 1969), pediments and alluvial fans (Weise 1974; Regard et al. 2006), as well as fluvial and marine terraces (Ehlers 1969, 1971; Grunert 1977). Recently, loess-soil sequences have been studied as well (Lateef 1988; Kehl et al. 2005a, 2005b; Frechen et al. in press). This paper briefly reviews this evidence in order to identify present knowledge gaps and research needs and to open new perspectives for future palaeoclimatological research in the area.

2 Modern climate

About 75% of the total land area of Iran is dominated by an arid or semiarid climate with annual precipitation rates from ~350 mm to less than 50 mm. In combination with potential evaporation rates, up to 3,000 mm a^{-1} high, the moisture deficit in the drylands is generally higher than 800 mm a^{-1} (Ehlers 1980). The dryness is caused by intense solar radiation and north-westerly to north-easterly winds, which mainly transport dry air-masses. It is further enhanced by the Alpine mountain ranges of the Alborz and Zagros (Fig. 1), preventing north-westerly and westerly depressions from the Caspian and Mediterranean Sea from entering the Iranian plateau. The convection of the moisture-laden air masses causes higher precipitation rates in the Caspian lowlands, the northern ranges of the Alborz and in the north-western part of the Zagros Mountains often exceeding 1,000 mm a^{-1}. Due to comparatively low evaporation rates because of higher cloud cover and altitude, the moisture surplus may exceed 1,200 mm a^{-1} in these areas (Ehlers 1980). Most of the precipitation falls during the months of October to April, except for the Caspian lowlands, where precipitation is more evenly distributed throughout the year (Fig. 1).

The mean annual precipitation rates and temperature data of selected climatological stations in Iran are shown in table 1. In the Caspian lowlands, a pronounced longitudinal precipitation gradient is reflected in mean annual precipitation rates decreasing from ~1,850 mm at Bandar Anzali in the West to ~435 mm at Gonbad-e Kavoos in the East. Precipitation rates in the Iranian highlands and the southern lowlands of the Khuzestan plain and the Persian Gulf coast are less than 350 mm a^{-1}.

In the Iranian highlands the mean annual temperature ranges from about 9 °C at Ardebil and Firouzkooh to about 22 °C at Zabol (Tab. 1). Lowest average daily temperatures between -3.9 °C and 8.6 °C are recorded in January, while the maximum average daily temperatures, up to 34.6 °C high, occur in July. The minimum and maximum monthly mean daily temperatures are higher in the southern lowlands, but the difference between the coldest and warmest months, i.e. the average amplitude of temperature is largest in the continental parts of the Highlands. Regional rainfall and temperature gradients are locally very pronounced, such as in the valleys of the Sefid-Rud or Zayandeh-Rud. These gradients are not reflected in the data. Hypsometric gradients of temperature and moisture are not well documented, due to a lack of climatological stations in the high mountains.

The general wind regime of Iran during the winter months is governed by air pressure gradients between the Siberian anticyclone and the equatorial low pressure system. In summer, a strong heat low

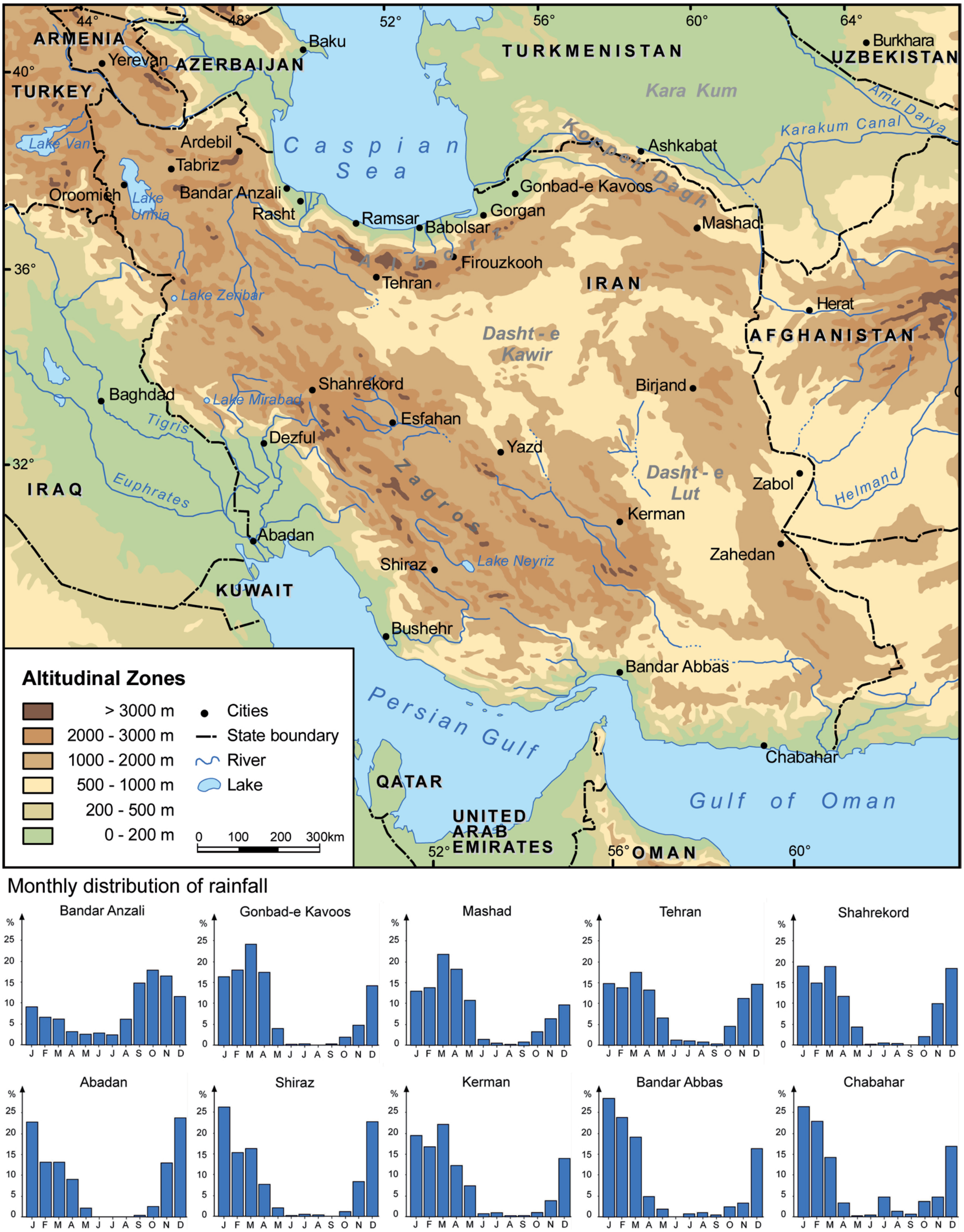

Fig. 1: Physiographic map of Iran and monthly precipitation for selected meteorological stations

Table 1: Precipitation and temperature data of selected meteorological stations in Iran (for locations see Fig. 1)

Station	Altitude	Mean annual precipitation	Monthly mean daily temperature (°C)			Mean annual temperature	Period
	(m a.s.l)	(mm)	Min.	Max.	Diffe-rence	(°C)	
South-Caspian lowlands							
Bandar Anzali	-26.2	1853.5	7.1[1)]	25.9	18.8	16.2	1951–2005
Ramsar	-20.0	1217.8	7.3[1)]	25.3[2)]	18.0	16.0	1955–2005
Babolsar	-21.0	894.4	7.8[1)]	26.5[2)]	18.7	17.0	1951–2005
Gorgan	13.3	601.0	7.9	27.8[2)]	19.9	17.8	1953–2005
Gonbad	37.2	435.8	7.8	29.9[2)]	22.1	18.5	1995–2003
Iranian highlands							
Oroomieh	1316	341.0	-1.8	23.9	25.7	11.5	1951–2005
Tabriz	1361	288.9	-1.7	26.0	27.7	12.5	1951–2005
Ardebil	1332	303.9	-2.5	18.3	20.8	9.0	1977–2005
Tehran	1191	230.5	3.7	30.2	26.5	17.2	1951–2003
Firouzkooh	1976	282.8	-3.9	20.5	24.4	8.8	1994–2003
Mashad	999	255.2	1.7	26.5	24.8	14.1	1951–2005
Shahrekord	2049	317.7	-1.5	24.0	25.5	11.8	1955–2003
Esfahan	1550	122.8	3.4	28.9	25.5	16.2	1951–2005
Shiraz	1484	346.0	6.2	29.2	23.0	17.7	1951–2005
Yazd	1237	60.8	5.9	31.9	26.0	19.1	1952–2005
Birjand	1491	170.8	4.4	27.7	23.3	16.5	1955–2005
Kerman	1753	152.9	4.6	26.7	22.1	15.8	1952–2005
Zabol	489	61.0	8.6	34.6	26.0	22.1	1963–2005
Zahedan	1370	90.6	7.3	28.5	21.2	18.4	1951–2005
Khuzestan plain and Persian Gulf area							
Dezful	143	404.6	11.5	36.3	24.8	24.0	1961–2005
Abadan	6.6	156.0	12.7	36.6	23.9	25.4	1951–2005
Bushehr	19.6	279.1	14.4	33.2[2)]	18.8	24.6	1951–2005
Bandar Abbas	10.0	185.5	17.8	34.3	16.5	27.0	1957–2003
Chabahar	8.0	113.9	19.9	31.4[3)]	11.5	26.2	1963–2003

1) coldest month in February, other stations in January

2) hottest month in August, other stations in July

3) hottest month in June

Data source: IRAN METEOROLOGICAL ORGANIZATION (2008)

develops over southern central Iran (GANJI 1968), and a relative pressure high prevails over Eurasia. This results in north-easterly to north-westerly winds blowing towards the Indian Ocean. Beginning in October and ending in April, westerly winds prevail, caused by depressions entering Iran from the Eastern Mediterranean along troughs of the 500 hPa flow patterns over Iran (ALIJANI 2002). Regional wind systems include the northerly to north-westerly "wind of the 120 days" during May to September in east-central Iran, the "shomal" ("north"), blowing along the northern coast of the Persian Gulf with peak intensities in June and July, and the "garmsil", a foehn-type wind which can occur in spring on the northern faces of the Alborz and Koppeh Dagh mountains (MIDDLETON 1986a). The strong winds often cause dust storms; their frequencies being exceptionally high in the Central Highlands, where 80.7 and 24.0 dust storm days per annum have been recorded at the cities of Zabol and Yazd, respectively (MIDDLETON 1986b).

Only the highest peaks of the Alborz and Zagros mountains such as the Kuh-e Damavand (5,671 m a.s.l.), Alam Kuh (4,850 m a.s.l.), Kuh-e Savalan (4,811 m a.s.l.) and Zardeh Kuh (4,548 m a.s.l.), bear small glaciers and exhibit features of active nivation and glaciation (BOBEK 1963). The total size of present-day glaciers amounts to about 20 km^2 (FERRIGNO 1988). In the southern Zagros and in south-eastern Iran the highest peaks with altitudes of more than 4,000 m apparently do not reach the present snowline, which can be explained by a dome-shaped deformation of the snowline above the dry central plateau (PREU 1984). However, the Shir Kuh (4,060 m a.s.l.), located south-west of the city of Yazd, has small permanent snow patches on its north-eastern slopes (GRUNERT et al. 1978). Unfortunately, more recent data on the extent of glaciers in Iran are not available in the scientific literature, and possible effects of global warming since the 1980-ies on glacier extensions have not been estimated.

3 Lower to Middle Pleistocene climate

For the Pliocene to Lower Pleistocene, slightly more humid climatic conditions than today have been postulated by Bobek (1963), who assumed that the brown silt and clay layers of the lower, 350 m thick section of the kavir fill at Masileh (Qom playa) described by Huber (1960) were deposited under a quasi-permanent lake environment. It is likely that during the same period, the Lut basin filled with horizontally bedded silty-clay siliclastics and intercalated evaporites, indicating a shallow and closed lake paleo-environment (Bobek 1969; Krinsley 1970; Dresch 1976). These sediments have subsequently been eroded by wind enforced by an increase in aridity to form the "giant Yardangs" of the Lut desert (Kavir-e Lut, Fig. 1).

The upper section of the kavir fill of Masileh was probably deposited from the Lower Pleistocene until the end of the last Glacial (Würm). It is composed of an alternating sequence of 4-5 salt banks and brown to green clay and silt layers with sand inclusions. The salt banks formed in playa environments under warm and dry climatic conditions, whereas the latter layers were interpreted as indicators of colder conditions with reduced evaporation and higher lake levels (Bobek 1963; Krinsley 1970).

Climate change during the Middle Pleistocene is also reflected in loess deposits of northern Iran (Kehl et al. 2005b; Frechen et al. in press; Fig. 2). In general, the accumulation of loess involves a series of climate-controlled processes including the production of mainly silt-sized particles, their deflation, eolian transport and deposition as well as syn- and post-depositional transformations including soil formation (e.g., Pécsi 1990; Gardner and Rendell 1994; Pécsi and Richter 1996; Pye 1995; Smalley 1995; Wright 2001). Unweathered loess can be correlated with stadial phases, reflecting dry (and cold) conditions during dust accumulation under a sparse vegetation cover, whereas paleosol horizons indicate comparatively moister (and warmer) conditions and steppe or forest vegetation during interstadials or interglacials (e.g.,

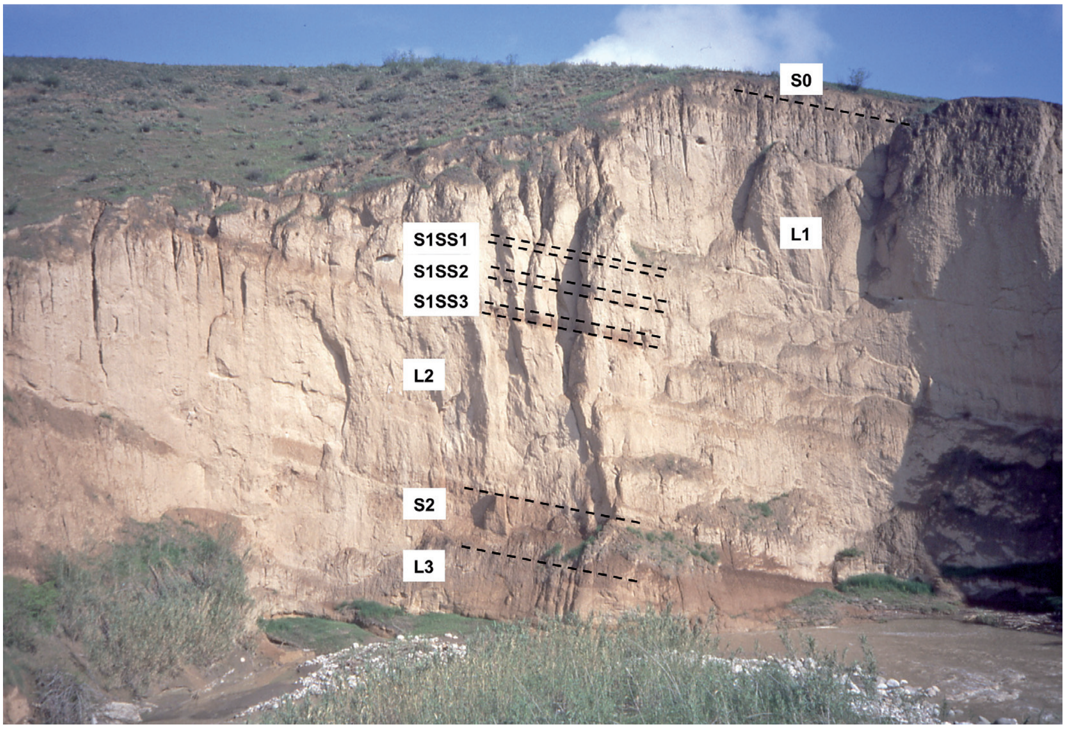

Fig. 2: The loess-soil sequence at Now Deh, northern Iran, about 30 m high, reflects climate changes between cold and dry glacials/stadials (unweathered loess L1, L2, L3) and warm and moist interglacials (brown soil horizons S1SS3, S2). The soil horizons S1SS1, S1SS2 and S1SS3 probably correlate with OIS 5a, 5c and 5e, respectively. In a river cut about 200 m downstream, S2 splits into three separate soil horizons, and two interstadial soils (L2SS1 and L2SS2), possibly formed during OIS 6, are exposed (cf. Fig. 3, lower part)

Fink and Kukla 1977; Dodonov 1991; Bronger 2003; Rutter et al. 2003). In the Sefid-Rud valley and on the northern foothills of the Alborz Mountains loess deposits contain several paleosols characterized by clay illuviation (Bt horizons). These paleosols developed under a forest vegetation and have tentatively been correlated with oxygen isotope stages (OIS) 7, 9 and 11 or older interglacials (Fig. 3). In addition, two weakly developed paleosol horizons (brown soil horizons of steppic soils) are intercalated with the loess deposit of OIS 6, giving the first evidence of Middle Pleistocene interstadials in Iran (Kehl et al. 2005b; Frechen et al. 2009). The chronostratigraphic estimates of paleosol formation in northern Iranian loess is based on physical dating of the sediment deposition ages using the luminescence method, and on counting the presumably interglacial soils "from the top". This implies a complete loess record, which is not necessarily valid. However, the earlier hypothesis of repeated Holocene cycles of loess accumulation and soil formation in Northern Iran (Bobek 1937, 1955; Barbier 1960; Ehlers 1971; Paluska and Degens 1980) must be rejected (Frechen et al. in press).

Recently, Djamali et al. (2008) have identified cyclical changes of pollen spectra in a 100 m long sediment core of Lake Urmia, a large (5.000 km²), hypersaline (> 200 g of salts / l) and shallow (8-12 m) lake located in north-western Iran (Fig. 1). The core sections with high percentages of arboreal pollen (AP) have been correlated with the OIS 7a, 5e, 5c and 5a, and those with very low percentages of AP and high percentages of *Artemisia* and Chenopodiaceae with stadial phases (Fig. 4). The correlation is based on two radiocarbon dates on

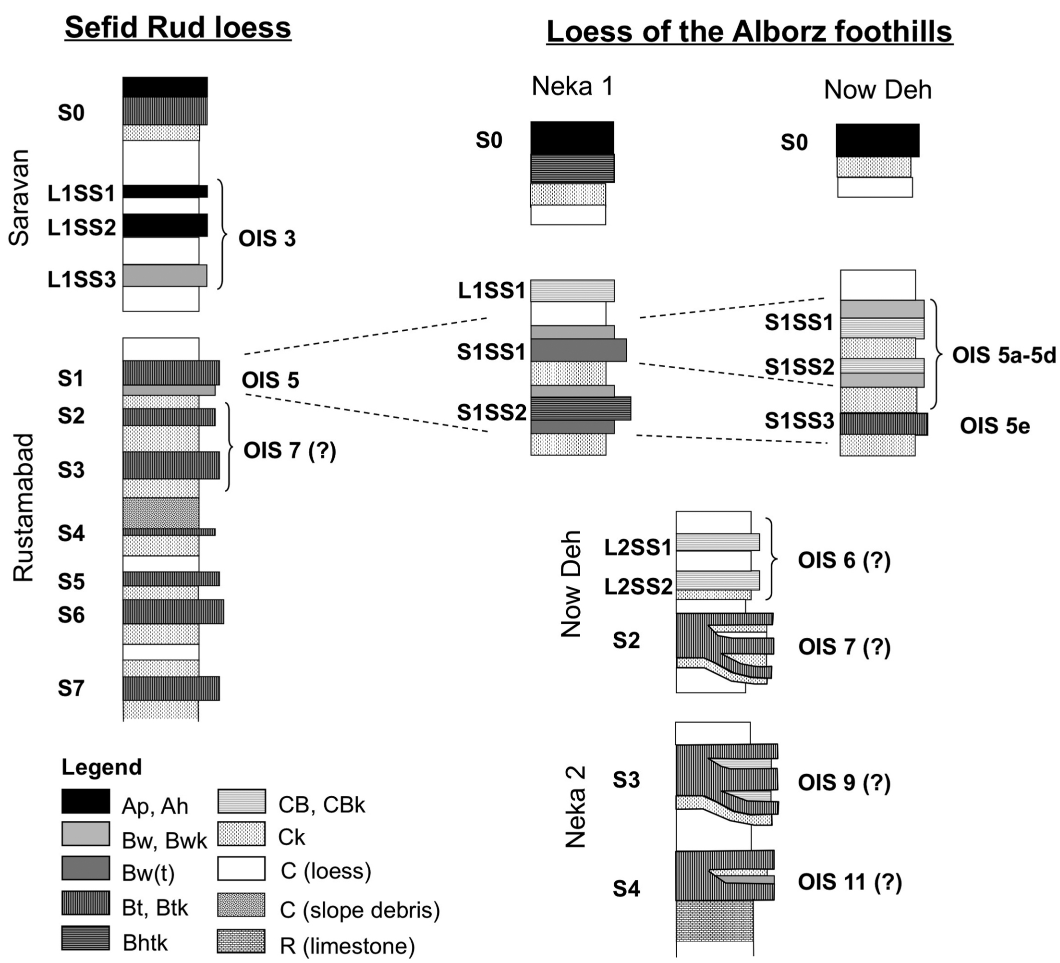

Fig. 3: Composite pedostratigraphies of northern Iranian loess in the Sefid-Rud valley (profiles at Saravan and at Rustamabad) and along the northern footslopes of the Alborz Mountains (profiles at Neka 1, 2 and at Now Deh) after Kehl et al. 2008, changed; denomination of soil horizons according to FAO (1998)

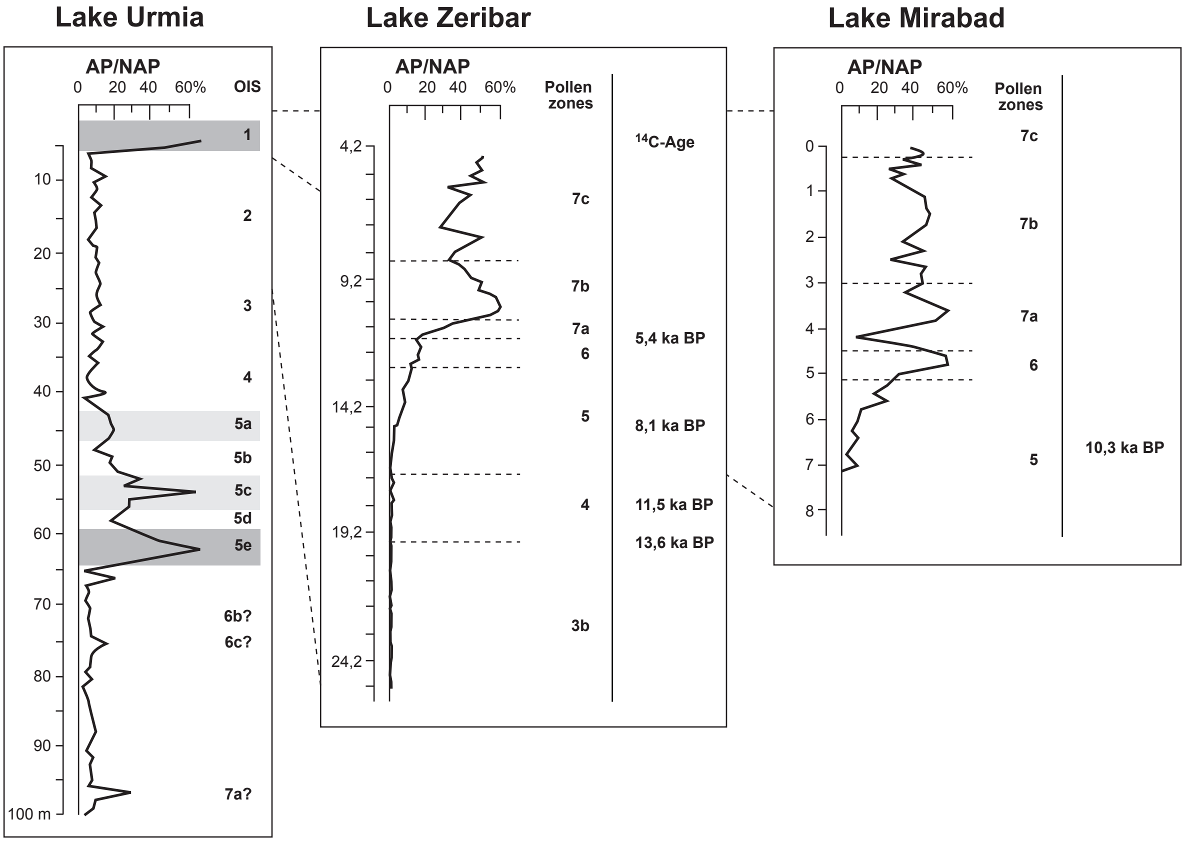

Fig. 4: Arboreal pollen, oxygen isotope stages (OIS), pollen zones (PZ) and radiocarbon dates of sediments from Lake Urmia (after Djamali et al. 2008, simplified) and lakes Zeribar and Mirabad (after Van Zeist and Bottema 1991, simplified)

bulk samples taken from 8- and 18.5-m depth yielding uncalibrated ages of 10345 ± 40 ^{14}C yr BP and 24,750 ± 200 ^{14}C yr BP (Djamali et al. 2008, 415) and on correlations with the Arabian Sea isotopic record (cf. Reichart et al. 1997) and a long pollen profile from Greece (cf. Tzedakis 1993, 1994). The pollen spectra of Lake Urmia show that during OIS 7a, or the Laylan Interstadial, according to the local chronostratigraphy of Djamali et al. (2008), a steppe-forest dominated by *Quercus* and *Juniperus* expanded near the lake, which was replaced by a steppe dominated by *Artemisia* and grasses during the penultimate glacial (Bonab Glacial). Shortly before the last interglacial an *Ephedra* shrub-steppe expanded near the lake, reflecting the climatic conditions of an interstadial (Djamali et al. 2008).

Stratigraphic evidence thus indicates Lower to Middle Pleistocene climate change in Iran, which might be reflected in geomorphic features, such as glacial moraines, fluvial terraces or lake terraces. For instance, based on the extent of playas, Krinsley (1970) postulated that the climate of the Iranian plateau was probably cooler and, because of reduced evaporation, moister than today. Several gravel terraces in the upper reaches of the Alborz and Zagros Mountains were correlated with the Lower to Middle Pleistocene and interpreted as indicators of cold phases with increased geli-solifluction (Bobek 1963). In the Alborz Mountains the rivers Sefid-Rud, Chalus, Haras, and Talar have formed several terrace levels situated at relative elevations of 70 to 150 above the present valley floor. These are (at least partly) composed of carbonate- or silica-cemented fluvial gravels and are possibly of pre-Würmian age (Ehlers 1969, 1971). Since increased tectonic uplift probably affected the formation of these terraces, their significance as evidence for climatic change is questionable. This may also hold for other geomorphic features including lake or marine terraces, alluvial fans and etchplanation surfaces, as long as numerical dating of the features is not available. Because of this uncertainty, Weise (1974) prefers not to correlate different phases of pedimentation observed in the Central Iranian Highlands with glacial or interglacial stages.

4 The Last Interglacial/Glacial cycle

The Lake Urmia core (Djamali et al. 2008) and the loess record of northern Iran (Kehl et al. 2005b; Frechen et al. 2009) probably cover large parts of the last climatic cycle. According to the pollen record of Lake Urmia the last interglacial (OIS 5e or Sahand Interglacial) was slightly warmer and moister than the Holocene. These climatic conditions were not reached during the interstadials 5c and 5a (Kaboudan Interstadials I and II), which are separated from each other and from the OIS 5e by two stadials (Espir Stadials I and II, Fig. 4).

In the loess record of northern Iran the last interglacial is represented by strongly developed Bt horizons of forest soils (Kehl et al. 2005b). It is likely that two moderately developed Bwk horizons of steppe soils in the loess section at Now Deh correlate with the interstadials 5c and 5a. However, based on preliminary luminescence age estimates, a correlation with OIS 3 would also be possible (Frechen et al. 2009). In the Sefid-Rud valley, two weakly developed Ah-horizons of steppe soils intercalated in loess at the Saravan section most likely correlate with interstadials of OIS 3. During stadials of the last glacial, dust accumulation reflects dry climatic conditions in which soil formation was hampered. The periods of dust accumulation can be correlated with the OIS 3 and OIS 2 up to the LGM, and in loess of the Iranian loess plateau also with the Lateglacial (Frechen et al. 2009).

Further stratigraphic evidence is found in the sedimentological record of Lake Zeribar, located in the western Zagros at an altitude of about 1,300 m a.s.l.. The dominance of herbaceous over arboreal pollen and a pollen assemblage indicating an *Artemisia* steppe reflect lower precipitation rates before and during the LGM than today (600-800 mm a^{-1}), the latter supporting the growth of a Zagros oak forest (Van Zeist and Wright 1963; Van Zeist and Bottema 1977; Fig. 4).

An alternative explanation for the dominance of *Artemisia* and lack of arboreal pollen has been seen in a relative increase in winter snowfall which might have suppressed the snow-sensitive *Pistacia* species (El-Moslimany 1986, 1987). The paleo-climatic deductions for the LGM based on pollen data have been partly corroborated by the content of plant macrofossils (Wasylikowa 1967, 2005; Wasylikova et al. 2006), *Cladocera* sp. (Megard 1967) and diatom assemblages of Lake Zeribar (Snyder et al. 2001). These data reflect lake water depth, water-table fluctuations and water salinity as a function of the lake water balance. Since the balance is controlled by a complex interplay of different factors, including precipitation, temperature-dependent evaporation, surface runoff, groundwater replenishment, and the elevation of the outlet (Stevens et al. 2001; Wasylikova 2005) it can only serve as an indirect estimate of the climatic conditions.

Recent investigations on a sand ramp near Ardakan in the Central Highlands (Thomas et al. 1997) and on loess deposits in southern Iran (Kehl et al. 2005a) substantiate the hypothesis of dry climatic conditions during the LGM which extended into the Lower Holocene. Several lines of evidence thus make it likely that the last glacial in northern Iran and the Zagros Mountains was a period of dry and more or less cold climatic conditions. Several estimates have been made on the temperature depressions during the Last Glacial including i) up to 5 °C in the Alborz and Zagros Mountains (Bobek 1963), ii) 5-8 °C during the LGM in central Iran (Krinsley 1970), or iii) between 8 and 10 °C for the mean annual temperature, and between 10 and 12 °C for the mean February temperature for southern Iran during the LGM (Frenzel et al. 1992).

In the southern Caspian lowlands about 25 km southwest of Babolsar (Fig. 1), a well developed buried paleosol derived from alluvial deposits has been reported by Antoine et al. (2006). Two radiocarbon ages indicate that the soil is younger than 28,486 ± 190 cal BP, older than 12,119 ± 82 cal BP, and thus could be correlated with the LGM or the Lateglacial. Since soil formation is promoted by precipitation, the paleosol is likely to reflect moist climatic conditions, whereas dust deposition during the LGM and the Lateglacial recorded in the loess deposits of Neka and Agh Band located about 75 km and 300 km to the east of Babolsar, respectively (Frechen et al. 2009), would represent rather dry conditions. This indicates past edaphic moisture gradients along the northern footslopes of the Alborz Mountains, a hypothesis which deserves further investigations.

Several studies have used geomorphological indicators as evidence of climate change during the Last Glacial. A series of glacial moraines in the Alborz and Zagros mountains testify to several glacier advances and recessional stages. Most moraines have been correlated with the Last Glacial implying several phases of snow line depression (Bobek 1963; Preu 1984). Main centers of former glaciation were the Takht-e Soleiman group including the Alam Kuh (e.g., Bobek 1937) in the Central Alborz, the Kuh-e Savalan (Schweizer 1970), the Irano-Turkish and adjoining Irano-Iraqi border ranges (e.g., Wright

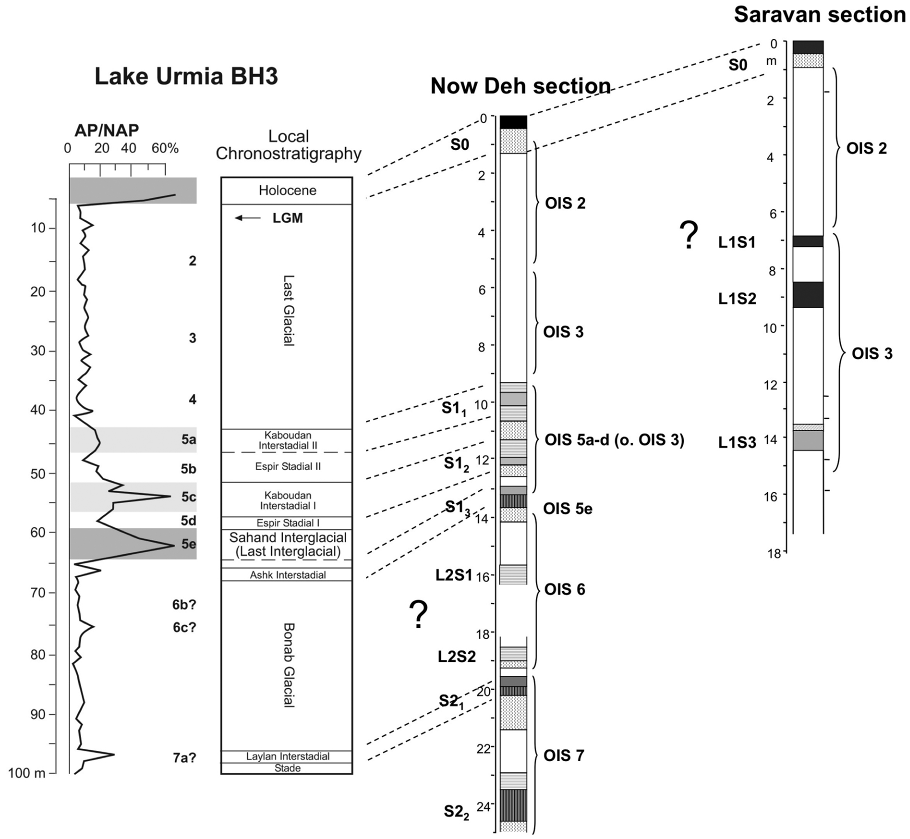

Fig. 5: Correlation of the pollen record from Lake Urmia (Djamali et al. 2008, simplified) and loess-soil sequences at Now Deh and Saravan (Kehl et al. 2005b; Kehl et al. 2008); legend see Fig. 3

1961; Schweizer 1975), and the Zardeh Kuh (e.g., Preu 1984) and possibly Kuh-e Dinar in the Central Zagros. The authors agree that the maximum extent of the glaciers in these areas was not much larger than at present, many of the former glaciers being confined to slopes and cirques. Further reports on former glaciation were given from the southern Central Highlands, an area which receives considerably less modern rainfall than the high mountains of Alborz and Zagros. Based on field observations of glacial moraines, scratchings on exposed bedrock, erratic blocks, and U-shaped valleys at the Shir Kuh and at the Kuh-e Jupar (4,135 m a.s.l.), Hagedorn et al. (1975) and Kuhle (1976) assume that large valley glaciers having a maximum ice thickness of 350 m (Kuhle 2008) existed and joined to form piedmont glaciers. Weise (1974) rejects to this hypothesis and suggests that part of the so-called glacial deposits in the high mountains south of Kerman is fluvially reworked rock-fall debris, and the apparent glacial cirques are scarps of similar shape. Here, further investigations including physical dating of exposure ages of bolders and scarps may shed light to this open question concerning the extent of Last Glacial glaciation in the southern Central Highlands.

Further geomorphic indicators of Last Glacial climate change were seen in lake terraces (Bobek 1937; Krinsley 1970; Schweizer 1975; Kelts and Shahrabi 1986) as well as in the extent of limnic to brackish deposits around Lake Urmia (Bobek 1963). It was estimated that during the Last Glacial Lake Urmia was raised to a maximum of 50–60 m (Bobek 1963) above its present water levels. However, Bobek (1963) concludes that the Lake Urmia terraces are not clearly related to past lake levels and that they apparently result from differential tectonic uplift. New evidence of relatively high lake levels in Lake Urmia during the last and penultimate glacial has been derived from pollen spectra including *Pediastrum*, dinoflagellate cysts and carbonate and organic matter deposition as indicators of brackish water conditions in Lake Urmia (Djamali et al. 2008). These brackish water conditions reflect increased runoff and/or decreased evaporation in contrast to the hypersaline conditions of today. According to Krinsley (1970), the surfaces of lakes Shiraz and

Neyriz in the southern Zagros Mountains were 2-3 m above their present playas during the Last Glacial Maximum (LGM).

The glacial high stands of Iranian lakes have been attributed either to increased precipitation, assuming the existence of pluvial periods or pluvial effects (Ehlers 1971), or to reduced evaporation of lake water related to cooler temperatures and accompanied by lower precipitation (Bobek 1937; Krinsley 1970; Stevens et al. 2001). During dry phases, the lower evapotranspiration of steppe-like vegetation, as compared to forest vegetation in humid phases, possibly increased stream-flow into the lakes and playas as suggested by Stevens et al. (2001) for Lake Zeribar. It is not unlikely that under more arid conditions and a less dense vegetation cover the relative proportion of surface runoff, compared to evapotranspiration and groundwater discharge, will increase, in particular if the higher aridity is accompanied by a shift to more sporadic high intensity rainfalls (Rohdenburg 1970). Thereby, the reduction in precipitation and total discharge may be more than offset by a gain in surface water discharge to lakes and playas.

The formation of fluvial terraces in the Alborz (Ehlers 1971; Paluska and Degens 1980) and Zagros mountains (Oberlander 1965; Fürst 1970) was possibly affected by tectonic uplift, which renders a climatogenetic interpretation and correlation with stadials of the Last Glacial very difficult. A similar problem may arise with explaining the formation of alluvial fan sequences. Recently, Regard et al. (2006) have reported alluvial fan abandonment ages determined by ^{10}Be exposure-dating in southern Iran of a) 44.0 ± 3.4 ^{10}Be ka, b) 20.1 ± 1.5 ^{10}Be ka, c) 12.8 ± 1.0 ^{10}Be ka, d) 8.4 ± 1.0 ^{10}Be ka and e) 5.6 ± 0.6 ^{10}Be ka. While abandonment ages a) and d) are explained by regional tectonic uplift the other three dates have been correlated with large scale climate events, i.e. with b) the end of the LGM, c) the onset of the Younger Dryas, and e) the Mid- to Late-Holocene climatic transition as evidenced in the Lake Zeribar cores. In the latter three cases a shift to drier climatic conditions is assumed to have been responsible for fan incision and abandonment of fan surfaces. This postulates that in southern Iran climate after the LGM became drier and that during the Lower Holocene precipitation was higher than today. As pointed out by the authors an indirect effect of climate on fan formation is the sea level lowering of about 100-120 m during the LGM, which exposed the sea floor of the Persian Gulf (e.g., Uchuqui et al. 1999) and changed the local base level of erosion.

5 Lateglacial and Holocene

The temperature and moisture conditions after the LGM have been well documented for the western Zagros Mountains in the sediment cores of Lake Zeribar and Lake Mirabad. As mentioned above, the LGM was characterized by dry steppe dominated by *Artemisia* plants (Van Zeist and Wright 1963; Van Zeist and Bottema 1991). At about 14 ka, temperatures probably increased, enabling a vegetation change to a pistachio-oak savanna, while precipitation remained low throughout the Lateglacial and Lower Holocene. Since about 6 ka, the temperature and rainfall characteristics have been similar to those of today, and Zagros oak forest grew in the catchments of Lake Zeribar and Lake Mirabad (Fig. 4).

The palaeolimnological records of the lakes (Griffith et al. 2001; Megard 1967; Snyder et al. 2001; Wasylikowa 1967, 2005; Wazylikowa et al. 2006) indicate water level fluctuations partly corroborating the palaeoclimatic conclusions deduced from the pollen record. However, each proxy record draws a slightly different picture, and the nature of climate change partly depends on the climate proxies considered (for discussion see Stevens et al. 2006 and Wazylikova et al. 2006).

The Lake Zeribar cores probably document the Younger Dryas climate episode. A significant increase in δ^{18}O (Stevens et al. 2001), maximum inferred salinity (Snyder et al. 2001, Wazylikova et al. 2006), and a pronounced phase of lake lowering between 12,600 and 12,000 cal BP, as indicated by plant macrofossils (Wazylikova 2005; Wazylikova et al. 2006) testify to a dry and cold period. However, the dry phase apparently did not lead to a change in vegetation cover (Bottema 1995). A concomitant dry spell may possibly be represented by the playa stage of Lake Urmia, inferred from sedimentological evidence by Kelts and Shahrabi (1986). In southern Iran, alluvial fan abandonment has been correlated with the Younger Dryas (Regard et al. 2006), as mentioned above.

The pollen profile of Lake Zeribar shows many similarities with the one of Lake Van, although in eastern Turkey the change to present-day climatic conditions probably started as early as 8.2 ka (Wick et al. 2003). In eastern Turkey and western Iran, forest invasion during the Lower Holocene was retarded, compared to coastal areas of the Near East, which experienced a period of relatively high precipitation during the Lateglacial and Lower Holocene (Goodfriend and Magaritz

1988; Roberts and Wright 1993). Comparatively dry climatic conditions possibly also prevailed in southern Iran, where the formation of the strongly developed Holocene climatic climax soil in the southern Zagros probably did not start before a shift to more humid conditions at 7 ka (Kehl et al. 2009). Possible causes for the Lower-Holocene dry period in the Taurus/Zagros mountain ranges are a strengthened outflow of cold air from the Eurasian landmass and a weakening of moisture supply by westerly depressions (Roberts and Wright 1993; Stevens et al. 2001).

After the Lower Holocene dry period, climatic changes probably occurred in Iran (Ganji 1978), but these are still poorly documented. Changes in seasonality of the precipitation and a short-lived increase of aridity between about 4 and 3.5 ka were inferred from stable isotope signatures of calcareous sediments of the Lake Zeribar and Lake Mirabad cores (Stevens et al. 2001; Stevens et al. 2006). Fluctuations in arboreal pollen spectra and clay layers in peat of a small mire located in the Central Caspian forest possibly report climate change at around 900 cal BP, and around 350 to 400 cal BP. However, these effects may have just as well resulted from human disturbance (Ramezani et al. 2008).

6 Pluvials or wet phases during the Late Quaternary?

Most evidence cited above points to the alternation of dry and cold climatic conditions during the stadials and comparatively moist and warm conditions during the interglacials and interstadials in Iran. Some authors postulate the occurrence of pluvials in Iran, assuming higher precipitation rates during the glacials than today. Scharlau (1958) hypothesizes that alluvial valley fill of the eastern Alborz Mountains was deposited under pluvial conditions during the last and penultimate Glacial. Sediments of a brackish lake, including marly peat, were found near the city of Kerman and interpreted as indicators of the Last Pluvial (Huckriede 1961). Vita-Finzi (1969) assumes that increased cyclonic precipitation during the Last Glacial (here 50 to 6 ka) and the Little Ice Age (A.D. 1550–1850), interrupted by a dry phase of erosion, caused the formation of two distinct alluvial deposits, which he defined as the Tehran and Khorramabad alluvium, respectively. A photograph of the latter (Fig. 10 in Vita-Finzi 1969) clearly shows several intercalated brown paleosol horizons, which possibly indicate climate-driven changes of sediment deposition and soil formation. Also, the formation of fluvial terraces in the Shir Kuh range and its postulated extensive glaciation were correlated with the Last Pluvial (Grunert 1977). Finally, Busche et al. (2002) relate stepped etchplains structuring the slope of the Kuh-e Namak salt dome near Ghom to pluvials of the Quaternary. The postulated pluvial conditions are often explained by a south-eastward shift of westerly depressions (e.g., Huckriede 1961; Ehlers 1971; Dresch 1976).

Possibly, part of the pluvial effect rather relates to wet phases, i.e. periods of increased precipitation during the Upper Pleistocene and Lower Holocene. Several Pleistocene wet phases occurred in the Eastern Mediterranean (Wagner and Geyh 1999), the south-eastern Sahara (Pachur and Hölzmann 1991; Hölzmann et al. 2004), and the Arabian Peninsula (McClure 1976; Nettleton and Chadwick 1996; Bray and Stokes 2004). A Pleistocene wet phase may also be indicated in Afghanistan (Pias 1972a, b). Lower-Holocene wet phases have been evidenced for the Arabian Peninsula in Yemen (Nettleton and Chadwick 1996), northern Oman (Burns et al. 1998; Fleitmann et al. 2003), the Rub Al Khali (Bray and Stokes 2004), and for western India (Roberts and Wright 1993), probably extending the humid belt of northern Africa (e.g., Pachur 2001; Hölzmann et al. 2004) to the east. These wet phases were probably caused by northward shifts of the Indian Monsoon (e.g., Sirocko et al. 1991; Hölzmann et al 2004; Staubwasser and Weiss 2006). Palaeopedological evidence suggests that Late Quaternary phases of increased humidity also occurred in presently semi-arid or arid parts of Iran. In Southern Iran, a well developed paleosol probably formed between 27 and 21 ka, as indicated by luminescence age estimates of the underlying und overlying loess-like sediments (Kehl et al. 2009), and argillic horizons in surface soils (Khormali et al. 2003) have been related to increased rainfall. Also, the stable isotope composition of pedogenic gypsum and carbonate and the occurrence of argillic horizons in polygenetic soils in Central Iran indicate phases of increased humidity (Khademi et al. 1997; Khademi and Mermut 1999; Farpoor et al. 2004). However, our estimates of the timing, intensity and duration of pedogenic processes are insufficient to decide whether the inferred increased humidity correlates with interstadials or with wet phases of the monsoon system.

7 Holocene climate change or human imprint?

Evidence of the timing and magnitude of climate events and anthropogenic imprints on landscape evolution in Iran is still sparse. In the western Zagros Mountains the domestication of goats probably began about 10 ka ago (Zeder and Hesse 2000). It is likely that with the onset of nomadic or peasant stock-breeding and cultivation during the Lower and Middle Holocene the ecologically sensitive drylands of Iran were affected by desertification, which nowadays is apparent in diverse processes of land degradation, including the degradation of the natural vegetation, soil erosion by wind and water, soil salinization and alkalization, depletion of groundwater resources or land subsidence. It has been hypothesized that, in combination with partial deforestation, a significant man-made degradation of the natural vegetation cover occurred as early as 4 ka BP (Bobek 1959), causing or accelerating soil erosion. The percentages of Gramineae and *Plantago lanceolata* pollen, often taken as disturbance indicators, strongly increase at about 4000 cal BP and again at 700 cal BP in the Lake Mirabad record (Stevens et al. 2006). As mentioned above, the clay layers and arboreal pollen spectra investigated by Ramezani et al. (2008) give evidence for human disturbance as well.

Archeological findings testify to the abandonment of settlements in the Iranian plateau, for which human-induced degradation of the environment has been made responsible (Meder 1979). However, the interrupted utilization or final abandonment of settlements may have been caused as well by climate change or political events, as suggested by Ehlers (1971) for the semiarid to subhumid Turkmen steppe in northern Iran. Evidence for the impact of Holocene dry spells, e.g. the 8.2 or 4.2 ka climate events, on the breakdown of agriculture-based societies in western Asia is increasing (see review of Staubwasser and Weiss 2006). However, in Iran, signals of Holocene climate events and anthropogenic imprints are still few and more information is needed on how and since when the land has been significantly degraded.

8 Past wind regimes

During the Pleistocene, northerly winds probably prevailed on the central Iranian Highlands, as suggested by the "giant yardangs" of the central parts of the Lut basin. These yardangs are up to 150 km long and are separated from each other by parallel corridors incised to a maximal depth of 200 m into silty material of Pliocene to Lower Pleistocene age. During the relatively dry and cold periods of the Pleistocene a strengthened Siberian anticyclone possibly led to the weakening and southward shift of the palaeo-monsoon, as evidenced for the period of the LGM (Sirocko et al. 1991). Coinciding with the weakening of the monsoon, cyclonic storm tracks were possibly displaced southwards. Furthermore, the westerly depressions may have been weaker (Stevens et al. 2001). The northerly shomal, blowing along the Persian Gulf, may have been active during the cold stages, too. It probably deflated sediments from the then-dry Persian Gulf and deposited them in the Rub Al Khali and further southwards onto the shelf of the Arabian Sea (Glennie et al. 2002). Northern Iran was probably dominated by northerly to north-easterly winds blowing from the Central Asian deserts as suggested by the delineation of dune fields in the Karakum desert (Létolle and Mainguet 1993), and by the spatial distribution of loess deposits along the Koppeh Dagh and in Northern Iran (Lateef 1988). Westerly depressions that today occasionally enter Northern Iran after crossing the Caspian Sea were possibly weaker. Because of these mechanisms, paleo-wind directions in Iran during the glacial periods were probably dominated by north-westerly to north-easterly equatorial currents.

Northward shifts of the monsoon system are indicated by, e.g., the sedimentary architecture of the Wahiba Sand Sea or the speleothem records in Oman (Preusser et al. 2002; Radies et al. 2004; Burns et al. 1998; Fleithmann et al. 2003). These shifts partly correlated with wet phases in the Arabian Desert and Oman, and possibly affected the circulation pattern in the southernmost part of Iran.

There is geomorphic evidence that, during the Last Glacial, central Iran experienced stronger winds than today. A sand ramp near Ardakan (Thomas et al. 1997) accumulated during the LGM, as determined by luminescence age estimates. Surprisingly, it was mainly deposited by winds blowing from the southeast, i.e. from the opposite direction than postulated above.

9 Concluding remarks

Past climate change in Iran has often been deduced from geomorphological observations, and in many cases it is not clear which role tectonics played in the formation of landforms or sediments taken as indicators of climate change. More reliable stratigraphic evidence has recently been extracted from lake sediments and loess deposits.

Overall, authors apparently agree that several Quaternary climatic changes occurred in Iran, also on the dry Iranian highlands. However, the timing and direction of changes have been a matter of dispute. Most authors argue for drier and colder climatic conditions during glacial stages than today, the latter more or less representing the conditions of older interglacial periods. Stratigraphical evidence of Pleistocene or Lower Holocene wet phases in Iran with increased rainfall compared to today, as described for the south-eastern Sahara and the Arabian Peninsula, is few. Also, climate episodes and events such as the cold spell of the Younger Dryas, have so far been identified in some places only. Middle and Upper Holocene climate changes in Iran and their possible effects on civilizations have scarcely been documented.

Geochronological data of the timing of Quaternary climate change in Iran are still sparse and no absolute age estimates have been published for the Lower to Middle Pleistocene so far. For the last Glacial and the Holocene, ^{14}C ages, luminescence age estimates, ^{10}Be exposure dates and archaeological evidence are available, which give first ideas of local Late Quaternary chronostratigraphies. More physical dating studies are needed to identify the timing of climate change in this large and ecologically diverse country.

Lake sediments in western Iran and loess-soil sequences in northern Iran have shown to be excellent archives of climate change, and further studies including high-resolution sampling of these records may yield more detailed information of past climates. The potential of other climate records, such as playa deposits of the Central Plateau or speleothems or tree-rings in the Zagros and Alborz Mountains has not yet been challenged.

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