Chapter 10
Current Continental Base-Levels Below Sea Level and Distant from the Sea

10.1 General

The following continental base-levels are distant from the sea and have, assumingly, no connection to the current marine base-levels. They are defined herein as base-levels below sea level since their lowermost terminal portion is below sea level.

10.2 The Great Artesian Basin and Lake Eyre Basin, Australia

The Great Artesian Basin (hereafter GAB) (Fig. 8.1) is one of the largest and oldest groundwater basins in the world occupying a big portion of central Australia (Fig. 10.1).

The basin covers a surface area of some 1.711 million km^2 (Mudd 2000). The climate is semi-arid to arid, with low precipitation, which becomes higher, averaging 600 mm/y, in the elevated bounding mountains (Herczeg et al. 1991). The evaporation in the discussed basin is quite high varying between 2,500 and 3,000 mm/y.

The basin was formed in Mesozoic times, and it contains a sequence of confined Triassic to Cretaceous, detrital aquifers and alternating aquicludes, that attains a maximum thickness of 3,000 m. The configuration of the basin is of a huge synclinorium, whereby the aquifers are exposed to recharge mainly in the eastern margins of the Great Dividing Range in Queensland, and to a lesser degree in the western margins. This configuration results in a convergent groundwater flow to the Lake Eyre Basin which occupies a big portion of the GAB and covers a surface area of some 1.14 million km^2. Lake Eyre in the southwestern part of the basin serves as a continental endorheic terminal lake and base-level (Fig. 10.1) (Herczeg et al. 1991; Mazor 1995; Mudd 2000). The northern part of the GAB drains to the north to the Gulf of Carpentaria.

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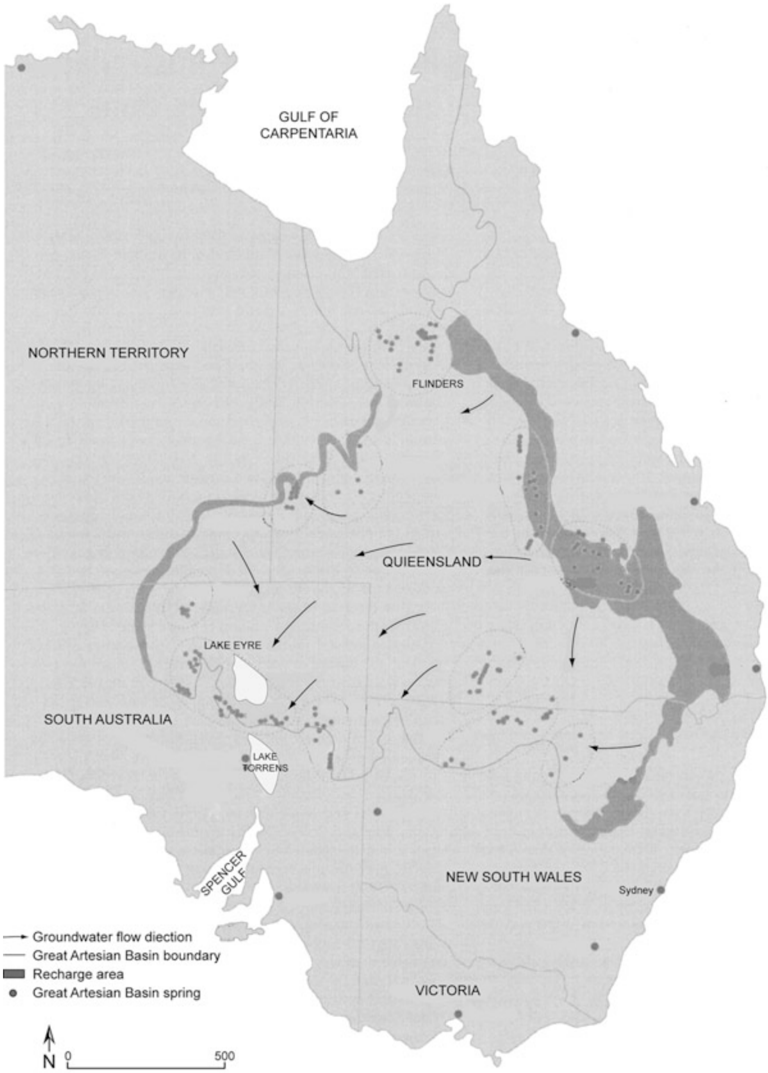

Fig. 10.1 A hydrological map of the Great Artesian Basin, Australia [after Mudd (2000)]

The aquifers of the GAB are confined and artesian, whereby their piezometric levels are above ground level over most of the GAB, excluding the margins of the basin.

In spite of being confined, the GAB is being discharged through vertical leakage to the regional water table and usually through hundreds of springs that emerge mostly close to the southern and southwestern margins of the basin (Figs. 10.1 and 10.2). Among these are the unique groundwater discharge features in the GAB of the artesian Mound Springs, that consist of carbonate deposited by the springs

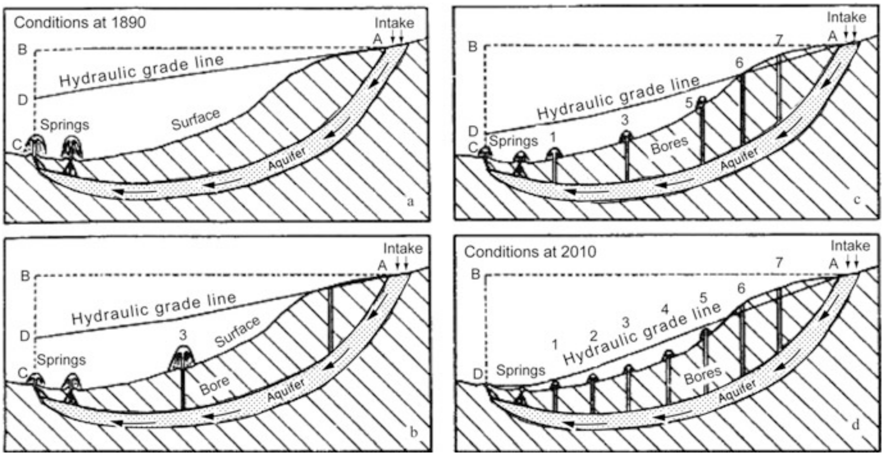

Fig. 10.2 Schematic hydrological sections in the Great Artesian Basin from the intake area to the discharge zone, showing the effect of groundwater exploitation on the decline of water heads and spring flows [after Mudd (2000)]

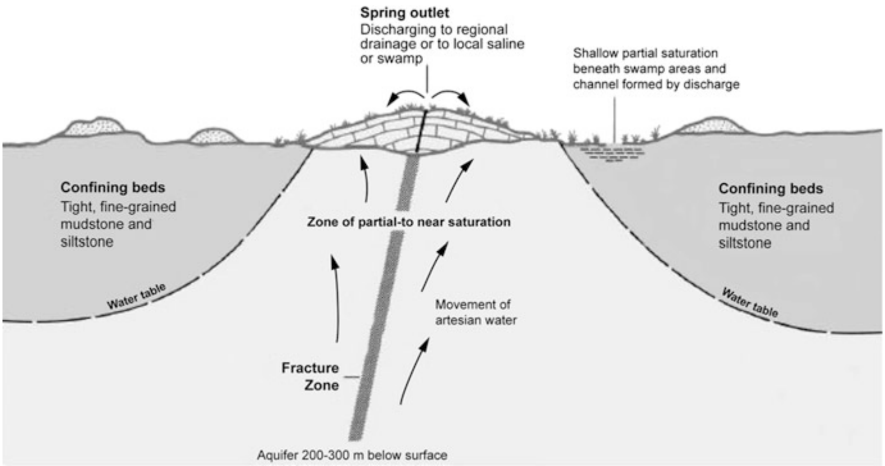

Fig. 10.3 The flow mechanism of a typical mound spring [after Mudd (2000)]

attaining, in places, an elevation of several meters (Fig. 10.3), as described in detail by Mudd (2000).

The age, as well as the residence and travel time of the groundwater of the GAB, from the intake areas to the discharge zones, were discussed in several articles. Mazor (1995) claimed that, due to the very slow groundwater flow in the basin, the aquifers are practically stagnant. Different from other studies, he attributed groundwater ages to be beyond the limits of the ^{36}Cl dating method, of one Ma. Namely, their ages are as old as 5–10 Ma or even more, based on the ^{4}He dating method.

Cresswell et al. (1999) described groundwater ages from the Amadeus Basin to be as old as 400 ka, and an average groundwater flow rate of 0.1 m/y, based on the ^{36}Cl dating method. Similarly, Collon et al. (2000) attributed to the GAB groundwater ages between 225 and 400 ka, based on the ^{81}Kr dating methods. In general, groundwater is relatively young, around several thousand years close to the intake areas and getting older along the flow path.

Groundwater salinity of the deeper Jurassic-early Cretaceous aquifer is between 500 and 1,000 mg/l, whereas the shallower Cretaceous aquifer already exhibits salinities of up to 10,000 mg/l (Mudd 2000). The geochemical characteristics of the GAB groundwater, as described by Herczeg et al. (1991), is as follows: The groundwater throughout the eastern and central parts of the basin is of the Na-HCO_3 type. Concentration of sodium and bicarbonate increases along the flow path toward the discharge zones in the west where the water changes to the Na-SO_4-Cl water type. The high Cl/Br weight ratio of these waters is indicative of evaporite dissolution.

Lake Eyre is the lowest terminal lake in the basin, at an elevation of 15 m bsl. The lake is practically an ephemeral huge playa or intermittent lake, formed by sporadic flooding, as happened several times during the last decades. Water salinity changes drastically in accordance to flooding and desiccation events. As an example, during one year, between 1984 and 1985 water salinity changed from 25 to 270 g/l (Williams and Kokkinn 1988). Evaporite beds and crusts are formed when water salinity reaches the saturation stage during the desiccation and dry time spans.

The paleolimnology and the paleo Lake Eyre level changes were discussed, among others, by Nanson et al. (1998), Magee et al. (2004) and DeVogel et al. (2004). According to these studies, Lake Eyre, which is at present an intermittent lake, was a perennial lake in the late Quaternary, during phases of increased monsoon activity, as evidenced by paleo shore lines that reflect paleo lake high-stands, related to the more humid phases. Those paleo shorelines were dated to be as follows (Fig. 10.4): Highstand of 10 m asl at about 120 ka, of 5 m asl at about 86 ka, of 3.5 bsl at about 63 ka and of 10 m bsl at about 40 ka. At the highest level at about 120 ka, the paleo perennial lake extended over a large area including the chain of the Eyre, Gregory, Blanche, Callabonna and Frome lakes.

The endorheic Lake Torrens (Fig. 10.1) is a 5,700 km^2 dry salt pan, a few tens of meter above sea level, south of Lake Eyre. Both lakes are separated by a topographic divide with a lowest elevation at 92 m asl indicating that paleo Lake Eyre could have never over-flown southward to Lake Torrens (DeVogel et al. 2004). The chloride content of the groundwater increases towards Lake Torrens and its brine enrichment occurs in the playa by evaporation from the capillary zone (Schmid 1988). It is important to note that Lake Eyre and Lake Torrens are aligned along a rift or a trough, which extends from the Spencer Gulf in the south, and is bounded to the east by the Flinders Ranges. Several studies showed that the area has been tectonically active, evidenced by subsidence of the rift, by young displacements, by current seismicity and by anomalous heat flow (Neumann et al.

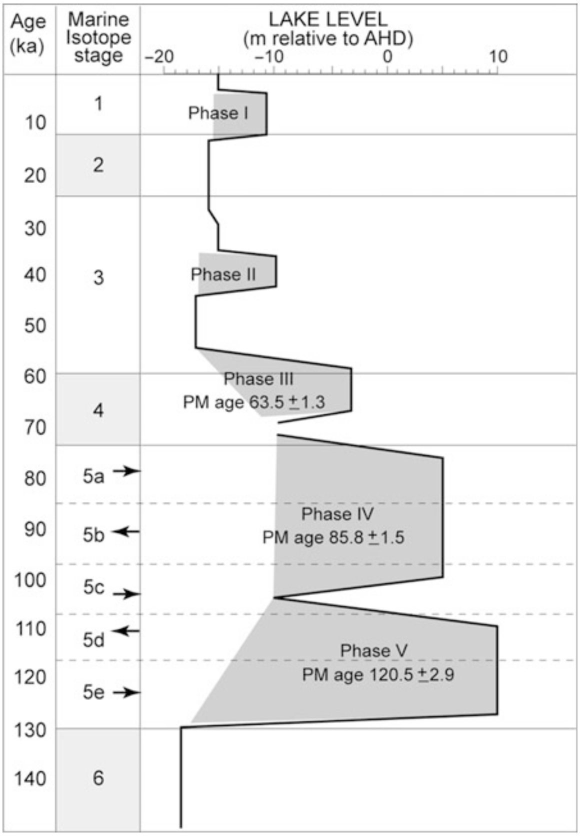

Fig. 10.4 Lake Eyre level curve for the past 150 ka [modified after Magee et al. (2004)]. PM ages denote pooled mean ages

2000; Sandiford 2003; Celerier et al. 2005; Quigley et al. 2006, 2007; Sandiford and Quigley 2009).

The Lake Eyre endorheic basin exhibits a combination of a continental base-level, only somewhat below sea level, located along a tectonically active trough which extends to the sea, and separated from the latter by an assumed rather low groundwater divide in between. Coupled with the existence of a geothermal anomaly it can be speculated that a subsurface seawater encroachment to the lower base-level may occur along the active faults. This process could be problematic in this case because of the long distance, and the relative, small elevation difference between Lake Eyre and the sea of only 15 m. The proposed additional mechanism of salination herein, by encroaching seawater, might be supported by the convecting seawater mechanism, tied to the existing geothermal anomaly, that was proposed for the Salton Trough by Barragan et al. (2001). The specific condition which favor salination due to seawater intrusion from the sea into low lying base level of the Salton Trough are elaborated in Sect. 11.5.

10.3 The Death Valley Basin

The Death Valley (Fig. 8.1) is a endorheic closed basin, at coordinates 36°N 117°W, in California, USA, attaining its lowest elevation at 86 m bsl. It is a north-south directed elongated valley, between the Amargosa Range (Black and Funeral Mountains) in the east and the Panamint range in the west (Fig. 10.5).

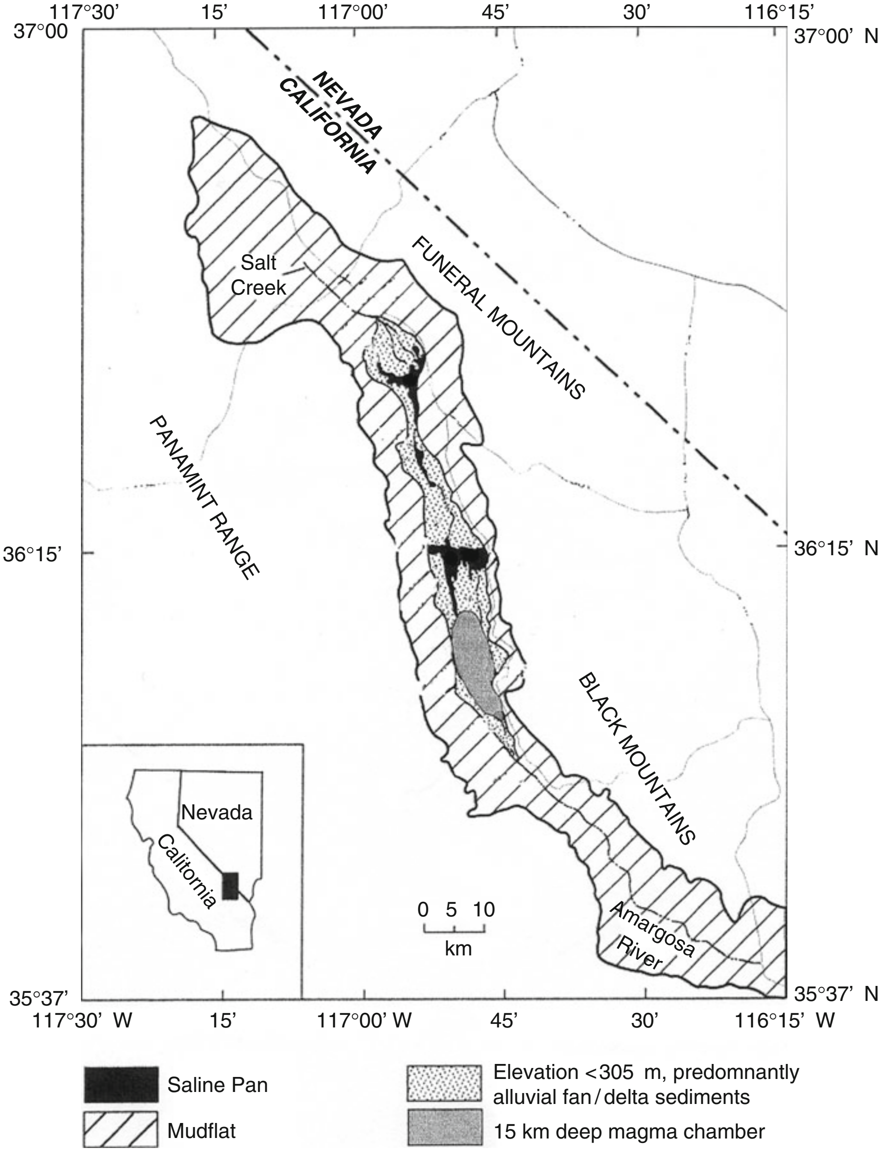

Fig. 10.5 Map of the Death Valley, California [after Lowenstein and Risacher (2009)]

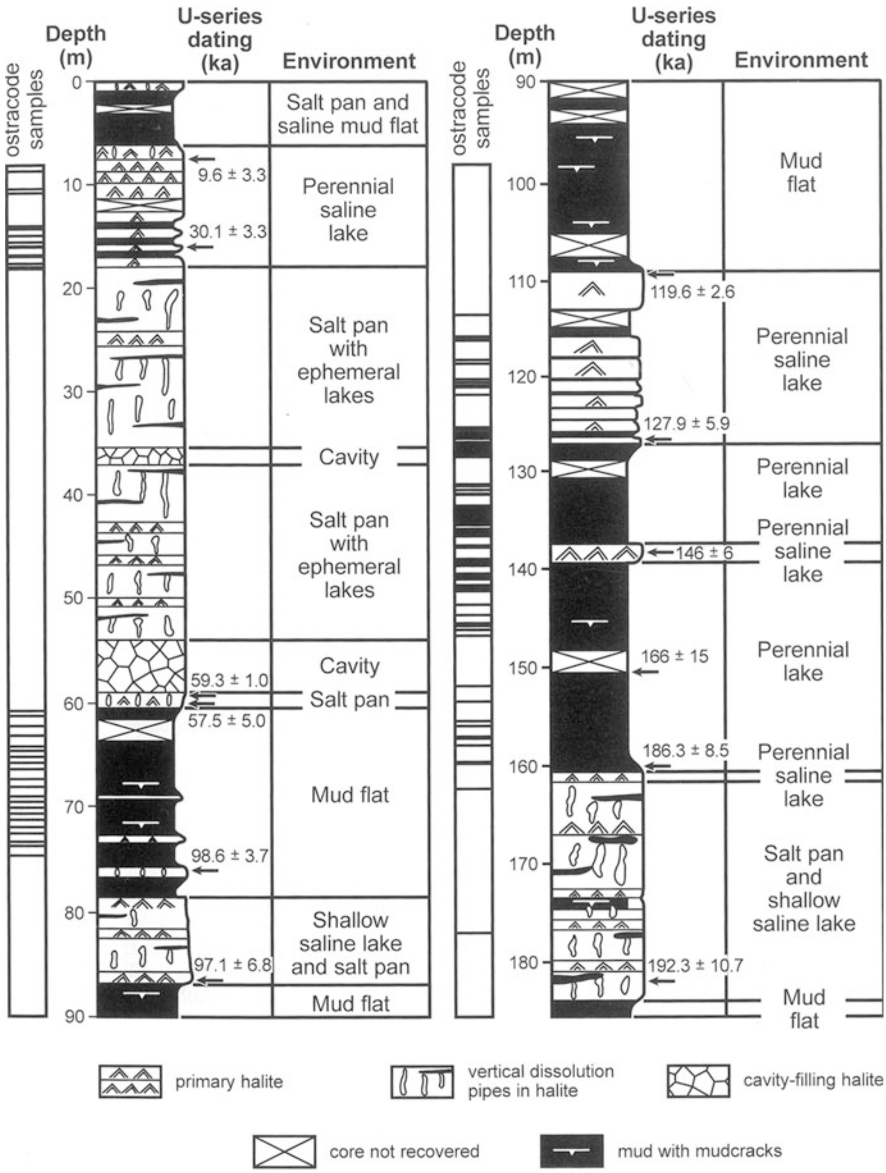

Fig. 10.6 A core columnar section in the Death Valley exhibiting the paleohydrological changes during the last 200 ka [after Lowenstein et al. (1999)]

Structurally speaking, the basin is part of the Basin Range province, formed as a pull apart basin and associated with strike slip and normal faults, which are still active (Lowenstein and Risacher 2009). The climate is arid, typified by an average precipitation of no more than 50 mm/y, and high temperatures and evaporation. The bottom of the valley consists of mud flats, salt pans and crusts (Fig. 10.5), as described by Crowley and Hook (1996) and Lowenstein et al. (1999).

The paleoclimate and paleohydrology of the basin were studied by several authors, among which are: Li et al. (1996, 1997), Ku et al. (1998), Lowenstein et al. (1999), Nelson et al. (2001), Forester et al. (2005) and Miner et al. (2007). The indicators that were used to reconstruct the paleohydrological regime of the basin were sedimentology, water salinity, stable and radiogenic isotopes of water, paleo spring deposits and ostracods. According to the above studies, the basin exhibits a closed basin characteristics in the last 200 ka, subjected to climatic changes and alternating hydrological regimes evidenced by the following: Dry periods, similar to the present one, are evidenced by mud flats, shallow saline lakes and salt pans which prevailed between 110 and 60 ka, and between 10 ka and the present. More humid periods are evidenced by perennial fresher lakes and perennial saline lakes, such as the occupied paleo Lake Manley between 185 and 128 ka and between 35 and 10 ka (Fig. 10.6).

The present day inflows to the Death Valley endorheic basin include rivers inflow from north and south and convergent groundwater and spring water flows that are recharged in the surrounding mountains intake areas (Fig. 10.7) (Bakker et al. 1999; Lowenstein et al. 1999; Anderson et al. 2006) .This pattern is supported mainly by the minimal direct recharge, related to the present arid climate, and the depleted stable isotope compositions of the groundwater, related to higher recharge altitudes (Larsen et al. 2001) However, Anderson et al. (2006) disagree with the

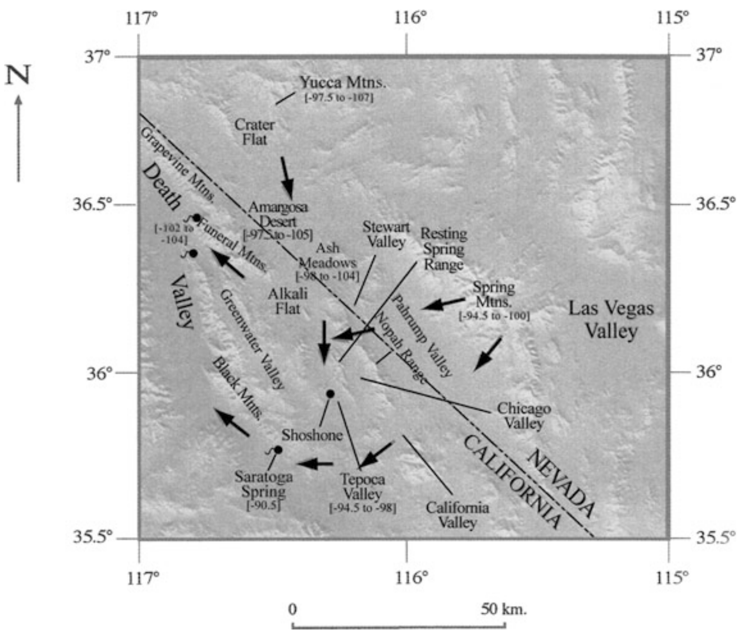

Fig. 10.7 Regional groundwater flow paths in the Death Valley area [after Nelson et al. (2001)]

conventional model of inter-basinal flow from distant parts of the basin, claiming that most of the recharge is local.

Higher salinities of groundwater are reached in the basin through interaction with saline lakes deposits (Larsen et al. 2001). An additional suggested saline contribution is the inflow to the basin of thermal waters from depth (Larsen et al. 2001), or more specific, hydrothermal Ca-Cl brines which emerge along active faults associated with an existing magma chamber at depth. This mechanism resembles the case of the Andean Altiplano and the Qaidam base-level in China, as described by Lowenstein and Risacher (2009) (see also Sects. 9.4 and 10.4).

The Death Valley base-level resembles, regarding its features and the hydrological regime, other endorheic continental base levels, described in this chapter.

10.4 Central Asia Closed Endorheic Basins

The vast region of central Asia (Fig. 8.1), which includes northwest China and Inner Mongolia, occupies hundreds of closed inland terminal river basins and lakes at elevations between 154 m bsl and thousands of meter above sea level. The basins are a few thousands kilometer distant from the nearest marine base level. These basins serve as base-levels to both surface and groundwater flows, occupying fresh as well as saline lakes, dry deserts and salt pans. This region is included technically, herein, under the category of base-levels below sea level due to its lowermost elevation, despite the higher elevations of most of its endorheic basins as described below.

The basins are, in general, inter-mountain basins situated structurally in between long mountain ranges such as the Altay, Tianshan, Qilian and Kulun mountains (Fig. 10.8). The entire region was subjected during its geological history to tectonic movements which culminated, induced by the Indian–Eurasian plates collision, since late Pliocene – early Pleistocene (Tapponnier and Molnar 1977). The mountain ranges were uplifted at a higher rate and the basins in between subsided, accumulating vast amounts of detrital basin fill sediments derived from the mountains, as well as lacustrine deposits (Edmunds et al. 2006; Zhu et al. 2007).

The paleo climatic and paleo hydrological regime of the region was reconstructed in several studies, based on paleo lakes shore lines, stable and radiogenic isotope compositions of the groundwater, sedimentological characteristics and pollen assemblages. According to these studies, the region was subjected to dry and humid climatic variations and thus to resultant different hydrological regimes since the late Pleistocene. The basins of northwestern China exhibit a warm and a more humid climate than the present one, during 40–30 ka BP (Yang et al. 2004). The region was also subjected to frequent climatic variations between humid and dry periods during the entire Holocene, between 10.7 and 1.5 ka BP (Hartmann and Wuenemann 2009), as is also evidenced by pollen assemblages (Feng et al. 2006).

Among the known closed basins of the region, those that are referred to below, are: The Tarim, Heihe, Minqin and Qaidam basins, all of which are, regarding their

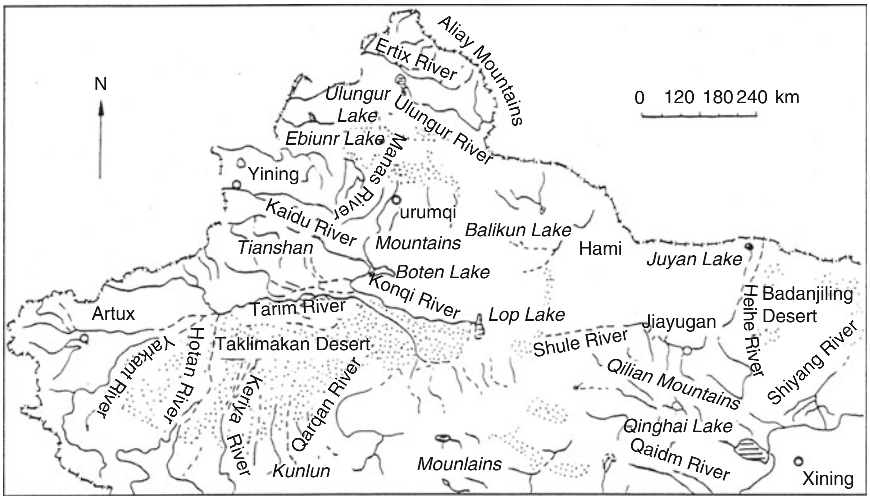

Fig. 10.8 Map of arid zones and stream systems in northwest China [after Wang and Cheng (2000)]

elevation, above sea level and the Turfan (Turpan) Basin which is considerably below sea level at its lower reaches. All these basins are characterized by an arid climate.

The Tarim Basin, is situated between the Tianshan Mountains in the north and the Kunlun Mountains in the south. Its terminal discharge zone is the Lop Nor Lake at coordinates 40°N 90°E and at the elevation of 768 m asl. The basin is drained by the largest inland river in China. The lower reaches of the basin and the lake are in the process of desiccation related to climate change and groundwater abstraction accompanied by water quality degradation. Groundwater salinity in the lower reaches of the basin rose to the level of 13 g/l (Feng et al. 2001, 2005).

The Heihe Basin extends from the Qilian Range in the south to its terminal Ejina Basin and Juyan lakes, at coordinates 41°N 100°E, in the lower reaches of the Heihe River. The Ejina Basin serves as a terminal discharge zone to both surface and groundwater flows. The convergent groundwater flow towards the Juyan lake in the north is displayed in the water table contour map of the basin (Fig. 10.9) (Wen et al. 2005). Groundwater salinity attained levels that exceed 5 g/l, in some places in the basin (Zhu et al. 2008). In the terminal northern part of the basin, the salinity is probably significantly higher as evidenced by the abundant salt crusts, as a result of super-saturation through evaporation (Wen et al. 2005).

The Minqin Basin, at coordinates 38°N 103°E, is situated at the lower reaches of the Shiyang River Basin, which extends from the Qilian Range in the south to the Minqin Basin in the north. The basin fill in the north is rich in evaporites such as halite and gypsum. The salinity of groundwater in the basin attained a level that exceeds 5 g/l. The basin was subjected to water table drop and water salination following severe over-exploitation (Zhu et al. 2007).

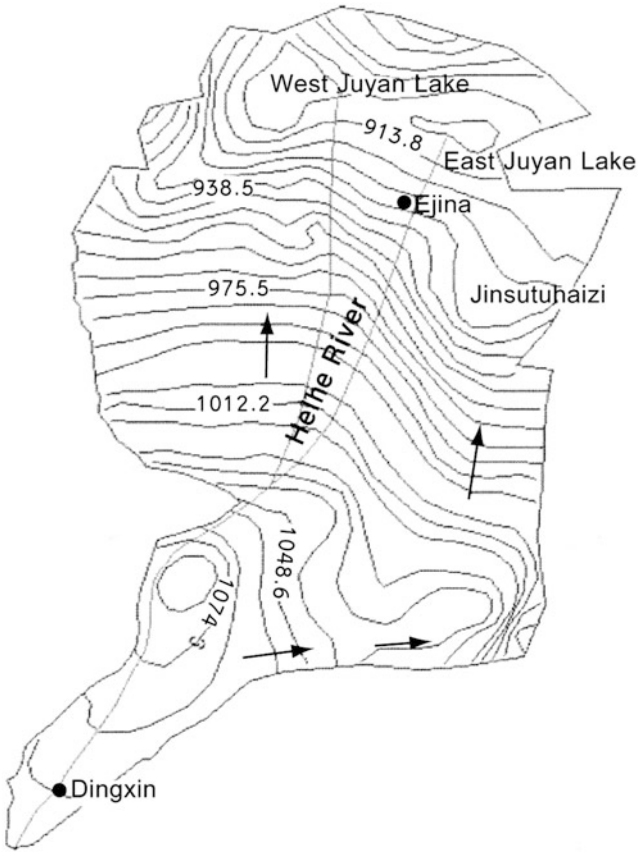

Fig. 10.9 Groundwater table map in the Ejina Basin exhibiting convergent flow toward the Juyan Lake system in the north [after Wen et al. (2005)]

The Qaidam Basin, at coordinates 37°N 94°E, is a terminal closed basin between the Kunlun Mountains in the south and the Qilian Mountains in the north. The center of the basin contains large areas of salt pans and saline lakes with the largest Qaharan Salt Lake plain. The Qaharan salt plain contains concentrated brines some of which are Ca-Cl brines (Lowenstein and Risacher 2009). Inflows to the center of the basin are from rivers that originate in the surrounding mountains, from converging groundwater flow from the mountain aquifers and from emerging hydrothermal Ca-Cl brines from depths, as found in other similar (Andean Altiplano, Death Valley) basins (Lowenstein and Risacher 2009) (see also Sects. 9.4 and 10.3).

The Turpan (Turfan) Basin is a terminal endorheic basin, at coordinates 42°N 90°E. The basin is structurally a fault bounded trough between the Tianshan Mountains in the north and the Quoltag Mountains in the south. The lowest point in the basin is the saline Lake Ayding at 154 m bsl that serves as the terminal base-level to surface and groundwater flows. The salinity of the lake is extremely high, exceeding 200 g/l. The inflows to the basin is only via rivers and groundwater that are fed by rainfall and snow melt and recharge that originate on the Tianshan Mountains in the north (Wei and Gasse 1999).

The present climate is generally arid in the lower parts of the basins and is more humid in the surrounding mountain ranges. As a result, the basins' aquifers are subjected to almost no direct recharge from rainfall in their lower reaches, but rather from the rivers that originate on the mountain ranges fed by rainfall as well as of snow and glacier melt and by lateral groundwater flow. Owing to the configuration of mountain ranges and closed basins, the rivers form a centripetal stream systems draining into the inland lakes or disappearing in the depression (Wang and Cheng 2000).

The basins' aquifers are recharged not only by groundwater convergent flows from the mountains aquifers (Edmunds et al. 2006) but also by the interconnected surface-groundwater systems as shown schematically in Fig. 10.10 (Wang and Cheng 2000; Ji et al. 2006).

A general down gradient geochemical evolution and zonation exist from the intake areas to the discharge zones, from fresh, mostly bicarbonate water types to farther more saline waters of bicarbonate–sulfate, sulfate–chloride and chloride water types (Ji et al. 2006). The terminal lower reaches of the basins occupy either brackish waters or hypersaline brines, attaining a salinity of up to 350 g/l in the Xingjiang lakes and up to almost 500 g/l of their adjoining interstitial waters (Zheng 1987). Additional potential salinity sources are emerging Ca-Cl brines, as described before from the Qaidam Basin by Lowenstein and Risacher (2009).

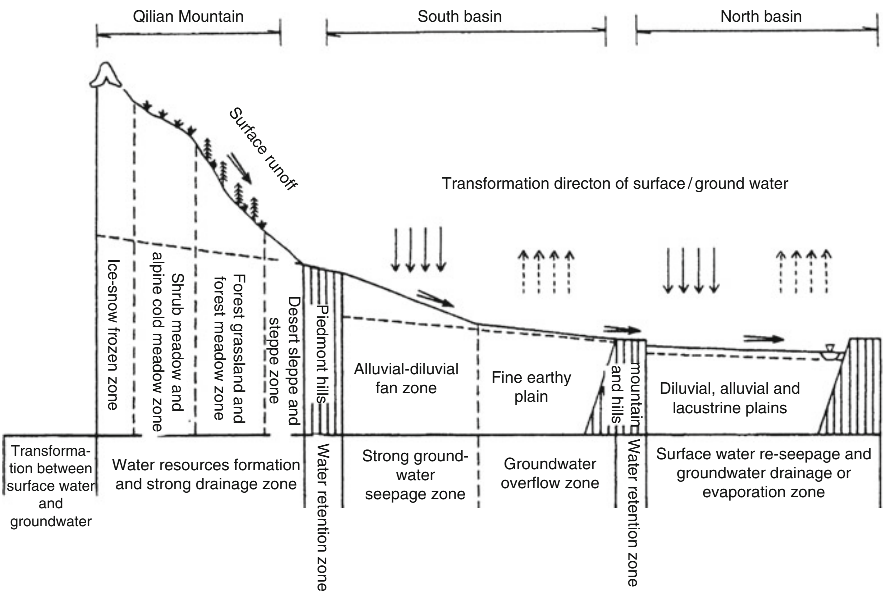

Fig. 10.10 A sketch of the hydrological regime of the different zones in the inland river basins of northwest China [after Wang and Cheng (2000)]

The geochemical evolution of groundwater in the discussed basins, along the flow path from the mountains intake areas to the terminal discharge zones, as well as the groundwater ages, were discussed in several studies (e.g., Zheng 1987; Feng et al. 2001, 2005; Wen et al. 2005; Edmunds et al. 2006; Ji et al. 2006; Chen et al. 2006; Zhu et al. 2007; Zhu et al. 2008; Lowenstein and Risacher 2009; and Su et al. 2009).

The hydrological characteristics, based on the above, that prevail in the discussed basins are as follows: The hydrological system consists, basically, of the mountain aquifers that are phreatic in the intake areas, and turn to be confined down-gradient within the lower parts of the basins. On top of that, a shallower phreatic aquifer system exists adjacent to the base levels.

Regarding stable isotope composition and groundwater age, it is noticed that the confined aquifers have a depleted stable isotope composition as compared to that of present day precipitation. This is in agreement with the tritium free content and the older ^{14}C ages of the groundwater, which is indicative of mostly paleo recharge in more humid periods. The shallow phreatic aquifers, on the other hand, contain tritium and are isotopically heavier than that of the present day precipitation, as a result of continuous evaporation along the flow path.

Regarding salinity and geochemical evolution of the groundwater, it is noticed that the shallow phreatic aquifers are more subjected to increase of salinity, as compared to the confined aquifers. The processes that are responsible for the salinity rise down-gradient are evaporation, water–rock interaction, dissolution of evaporites, over-exploitation, and to anthropogenic pollution.

10.5 The Caspian Sea-Kara Bogaz Gol Basin

The Caspian Sea basin (Fig. 8.1) is the largest closed terminal water body on earth, with a surface area of ca 380,000 km^2, located at coordinates 40°N 51°E (Fig. 10.11). Its present elevation is around 28 m bsl and it is fed mainly by the inflows of the Volga River which contributes over 80% of its water input, and the rest is contributed by the Kura and Ural rivers. Regarding the history of the basin, it started as part of the initial giant regional Sarmatian Sea in the Miocene around 17 Ma, which was later on, around 6 Ma, separated into the Caspian Sea, the Black Sea and Panonian basins. The Caspian Sea was later separated from the Black Sea at around 3 Ma (Klige and Myagkov 1992). Different from other authors, it was claimed by Dumont (1998) that the Caspian Sea was never connected to the Tethys Sea and should thus be regarded and termed as a lake.

In sub-recent historical times, the sea level fluctuated by some 6 m, but during its older geological history, fluctuations were over 200 m and the sea remained below sea level since the last glacial period (Dumont 1998). The level of the Caspian Sea fluctuated throughout the Pleistocene (Dumont 1998; Zubakov 2001) and throughout the last 2,500 years, but did not decline below 25 m bsl (Rychagov 1997). A dramatic water level decline of some 3 m was recorded during the 1930s (Giralt et al. 2003).

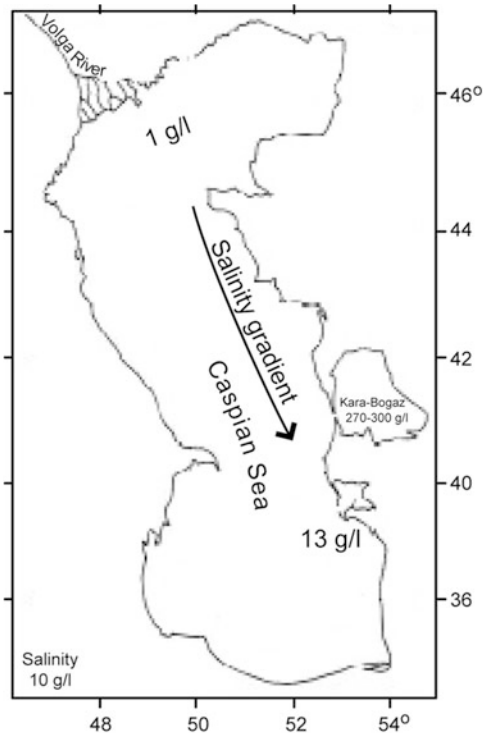

Fig. 10.11 Map of the Caspian Sea-Kara Bogaz system [modified after Dumont (1998)]. Also shown is the water salinity at various parts of the system

The hydrology and water chemistry of the present day Caspian Sea was studied, among others, by Dumont (1998), Clauer et al. (2000, 2009), Giralt et al. (2003) and Tuzhilkin et al. (2005). The Caspian Sea, at present, is an elongated water body, subdivided into three sub-basins (Giralt et al. 2003). It is shallow in the north, opposite the inlet and the delta of the Volga River and deeper in its central and southern portions. The vast amount of inflow from the Volga River to the shallow northern part of the sea, coupled with a more humid and cooler climate, as compared to the more arid southern parts of the sea, form a water salinity gradient to the south (Fig. 10.11). At the northwest, adjacent to the delta of the Volga River salinities are very low, whereas in the southeast salinity rises to 13 g/l, close to a third of ocean waters (Dumont 1998). The sea displays a significant multi-annual and spatial hydrochemical variability as reflected by the total water salinity of 7.2, 13 and 13.1 g/kg of the northern, central and southern sub-basins, respectively (Tuzhilkin et al. 2005).

The salinity rise toward the south occurs mostly through evaporation, but it was suggested, based on certain extra ion enrichment and specific isotope ratios (i.e., $^{87}Sr/^{86}Sr$), that an additional subterranean emerging hydrothermal water contribution also exists. This water leaches, while emerging, sedimentary evaporites, and is partly responsible for the specific water chemistry of the sea water. It also contributes to the rise of the sea level since about 1978 (Clauer et al. 2000, 2009). The main evaporite deposits are sulfates, as expected considering the water salinity.

The Kara Bogaz Gol Lake (KBG), east of the Caspian Sea, at coordinates 41°N 54°E (Fig. 10.11), is a shallow lagoon, a few meter deep with a surface area of some 18,000 km^2, located in an area of extremely arid climate. The hydrology of the KGB was discussed, among others, by Giralt et al. (2003), Leroy et al. (2006) and Kosarev et al. (2009). The KBG water level is several tens of centimeter to a few meter lower than that of the Caspian Sea and both are connected by a narrow strait, 110–300 m wide and 10–12 km long, through where the Caspian Sea overflows to the KBG. Following a considerable drop of the Caspian Sea level since the late 1970s, the straits were dammed, in March 1980, in order to stop this drop, and thus the inflow from the Caspian Sea to the KBG ceased. Following the closure of the straits, a quick desiccation of the KBG began, until it turned to a dry salt pan at the end of 1984. In 1992 the inflow from the Caspian Sea was renewed following the destruction of the dam.

The water salinity evolution of the KBG is controlled by the amount of Caspian Sea water inflow. Until the 1930s, the salinity was dominated by the vast influx of Caspian Sea waters of relatively moderate salinity and, thus, mostly sulfates (gypsum) were deposited at the bottom of the lagoon. Already in 1939 the water salinity increased, dominated mostly by a NaCl brine and halite started to be deposited since then. Following the drop of the KBG level in the 1970s, the water salinity rose to 27–30% and following the closure and disconnection from the Caspian Sea the brine salinity attained a concentration of up to 38%, accompanied by salt deposition of halite, epsomite, bischofite and carnallite.

The Caspian Sea is conventionally regarded everywhere as a terminal base-level. This statement is true when the entire basin, including the KBG is concerned. It is suggested herein that the Caspian Sea, proper, is in fact not naturally terminal by definition, since it overflows to the KBG and, thus, can be determined as a flow-through basin. As a result of this setup, the water salinity of the Caspian Sea remains intermediate, able to deposit sulfates only, such as gypsum, and not reaching salinities which are supersaturated with regard to halite. However, the KBG lagoon, downstream, is indeed the terminal part of the basin whereby salinities due to evaporation exceed the saturation threshold for halite and post-halite salts.

The Karagiye Depression, at coordinates 43°N 51°E, east of the Caspian Sea is a depression attaining an elevation as low as 132 m bsl. The entire area is described as saline lowlands. In spite of the scarce information, one can assume that this depression can potentially serve as an additional base-level to eastward density-driven brines from the Caspian Basin. Another depression, more to the east, at coordinates 41°N 58°E, namely the Akdzhakaya Depression, attains an elevation of 81 m bsl and is also described as a saline lowland. One can speculate that this depression functions, regarding salination processes, similar to the Karagiye Depression.

References

Anderson K, Nelson S, Mayo A, Tingey D (2006) Interbasin flow revisited: the contribution of local recharge to high-discharge springs, Death Valley, CA. J Hydrol 323:276–302

Bakker M, Anderson EI, Olsthoorn TN, Strack ODL (1999) Regional groundwater modeling of the Yucca Mountains site using analytic elements. J Hydrol 226:167–178

Barragan RRM, Birkle P, Portugal ME, Arellano GVM, Alvarez R (2001) Geochemical survey of medium temperature geothermal resources from the Baja California Peninsula and Sonora, Mexico. J Volcanol Geotherm Res 110:101–119

Celerier J, Sandiford M, Hansen DL, Quigly M (2005) Modes of active intraplate deformation, flinders ranges, Australia. Tectonics 24:TC6006:1–17

Chen Z, Nie Z, Zhang G (2006) Environmental isotopic study on the recharge and residence time of groundwater in the Heihe River Basin, northwestern China. Hydrogeol J 14:1635–1651

Clauer N, Chaudhuri S, Toulkeridis T, Blanc G (2000) Fluctuations of Caspian Sea level: beyond climatic variations?. Geology 28:1015–1018

Clauer N, Pierret M-C, Chaudhuri S (2009) Role of subsurface brines in salt balance: the case study of the Caspian Sea and Kara Bogaz Bay. Aquat Geochem 15:237–261

Collon P, Kutschera W, Loosli HH, Lehmann BE, Purtschert R, Love A, Sampson L, Anthony D, Cole D, Davids B, Morrissey DJ, Sherrill BM, Steiner M, Pardo RC, Paul M (2000) ^{81}Kr in the Great Artesian Basin, Australia: a new method for dating very old groundwater. Earth Planet Sci Lett 182:103–113

Cresswell RG, Jacobson G, Wischusen J, Fifield LK (1999) Ancient groundwaters in the Amadeus Basin, Central Australia: evidence from the radio-isotope ^{36}Cl. J Hydrol 223:212–220

Crowley JK, Hook SJ (1996) Mapping playa evaporite minerals and associated sediments in Death Valley, California, with multispectral thermal infrared images. J Geophys Res 101:643–660

DeVogel SB, Magee JW, Manley WF, Miller GH (2004) A GIS-based reconstruction of late quaternary paleohydrology: Lake Eyre, arid central Australia. Palaeogeogr Palaeoclimatol Palaeoecol 204:1–13

Dumont HJ (1998) The Caspian Lake: history, biota, structure, and function. Limnol Oceanogr 43:44–52

Edmunds WM, Ma J, Aeschbach-Hertig W, Kipfer R, Darbyshire DPF (2006) Groundwater recharge history and hydrogeochemical evolution in the Minqin Basin, North West China. Appl Geochem 21:2148–2170

Feng Q, Endo KN, Cheng GD (2001) Towards sustainable development of the environmentally degraded arid rivers of China – a case study from Tarim River. Environ Geol 41:229–238

Feng Q, Liu W, Si J, Su Y, Zhang Y, Cang Z, Xi H (2005) Environmental effects of water resource development and use in the Tarim River basin of northwestern China. Environ Geol 48:202–210

Feng X, Yan S, Ni J, Kong Z, Yang Z (2006) Environmental changes and lake level fluctuation recorded by lakes on the plain in northern Xinjiang during the late Holocene. Chin Sci Bull 51 Suppl 1:60–67

Forester MR, Lowenstein TK, Spencer RJ (2005) An ostracod based paleolimnologic and paleo-hydrologic history of Death Valley: 200 to 0 ka. Geol Soc Am Bull 117:1379–1386

Giralt S, Julia R, Leroy S, Gasse F (2003) Cyclic water level oscillations of the KaraBogazGol-Caspian Sea system. Earth Planet Sci Lett 212:225–239

Hartmann K, Wuenemann B (2009) Hydrological changes and Holocene climate variations in NW China, inferred from lake sediments of Juyanze palaeolake by factor analyses. Quat Int 194:28–44

Herczeg AL, Torgersen T, Chivas AR, Habermehl MA (1991) Geochemistry of ground waters from Great Artesian Basin, Australia. J Hydrol 126:225–245

Ji X, Kang E, Chen R, Zhao W, Zhang Z, Jin B (2006) The impact of the development of water resources on environment in arid inland river basins of Hexi region, Northwestern China. Environ Geol 50:793–801

Klige RK, Myagkov MS (1992) Changes in the water regime of the Caspian Sea. Geo J 27: 299–307

Kosarev AN, Kostianoy AG, Zonn IS (2009) Kara-Bogaz-Gol Bay: physical and chemical evolution. Aquat Geochem 15:223–236

Ku T-L, Luo S, Lowenstein TK, Li J, Spencer RJ (1998) U-series chronology of lacustrine deposits in Death Valley, California. Quat Res 50:261–275

Larsen D, Swihart GH, Xiao Y (2001) Hydrochemistry and isotope composition of springs in the Tecopa basin, southeastern California, USA. Chem Geol 179:17–35

Leroy SAG, Marret F, Giralt S, Bulatov SA (2006) Natural and anthropogenic rapid changes in the Kara-Bogaz Gol over the last two centuries reconstructed from palynological analyses and a comparison to instrumental records. Quat Int 150:52–70

Li J, Lowenstein TK, Blackburn IR (1997) Response of evaporate mineralogy to inflow water sources and climate during the past 100 k.y. in Death Valley, California. Geol Soc Am Bull 109:1361–1371

Li J, Lowenstein TK, Brown CB, Ku T-L, Luo S (1996) A 100 ka record of water tables and paleoclimates from salt cores, Death Valley, California. Palaeogeogr Palaeoclimatol Palaeoecol 123:179–203

Lowenstein KT, Li J, Brown C, Roberts SM, Ku T-L, Luo S, Yang W (1999) 200 k.y. paleoclimate record from Death Valley salt core. Geology 27:3–6

Lowenstein TK, Risacher F (2009) Closed basin brine evolution and the influence of Ca-Cl inflow waters: Death Valley and Bristol Dry Lake California, Qaidam basin, China, and Salar de Atacama, Chile. Aquat Geochem 15:71–94

Magee JW, Miller GH, Spooner NA, Questiaux D (2004) Continuous 150 k.y. monsoon record from Lake Eyre, Australia: insolation-forcing implications and unexpected Holocene failure. Geology 32:885–888

Mazor E (1995) Stagnant aquifer concept Part 1. Large-scale artesian systems-Great Artesian Basin, Australia. J Hydrol 173:219–240

Miner RE, Nelson ST, Tingey DG, Murrell MT (2007) Using fossil spring deposits in the Death Valley region, USA to evaluate palaeoflowpaths. J Quat Sci 22:373–386

Mudd GM (2000) Mound springs of the Great Artesian Basin in South Australia: a case study from Olympic Dam. Environ Geol 39:463–476

Nanson GC, Callen RA, Price DM (1998) Hydroclimatic interpretation of quaternary shorelines on South Australian playas. Palaeogeogr Palaeoclimatol Palaeoecol 144:281–305

Nelson ST, Karlsson HR, Paces JB, Tingey DG, Ward S, Peters MT (2001) Paleohydrologic record of spring deposits in and around Pleistocene pluvial Lake Tecopa, southeastern California. Geol Soc Am Bull 113:659–670

Neumann N, Sandiford M, Foden J (2000) Regional geochemistry and continental heat flow: implications for the origin of the South Australian heat flow anomaly. Earth Planet Sci Lett 183:107–120

Quigley M, Sandiford M, Fifield K, Alimanovic A (2006) Bedrock erosion and relief production in the northern flinders ranges, Australia. Earth Surf Process Landf 32:929–944

Quigley M, Sandiford M, Cupper ML (2007) Distinguishing tectonic from climatic controls on range-front sedimentation. Basin Res 19:491–505

Rychagov GI (1997) Holocene oscillations of the Caspian Sea, and forecasts based on palaeogeographical reconstructions. Quat Int 41–42:167–172

Sandiford M (2003) Neotectonics of southeastern Australia: linking the quaternary faulting record with seismicity and in situ stress. In: Hillis RR, Muller D (eds) Evolution and dynamics of the Australian Plate. Geological Society of Australia Special Publication, Australia 22:101–113

Sandiford M, Quigley M (2009) TOPO-OZ: insights into the various modes of intraplate deformation in the Australian continent. Tectonophysics 474:405–416

Schmid RM (1988) Lake Torrens brine. Hydrobiology 158:267–269

Su YH, Zhu GF, Feng Q, Li ZZ, Zhang FP (2009) Environmental isotopic and hydrochemical study of groundwater in the Ejina Basin, northwest China. Environ Geol 58:601–614

Tapponnier P, Molnar P (1977) Active faulting and tectonics in China. J Geophys Res 82:2905–2930

Tuzhilkin VS, Katunin DN, Nalbandov YR (2005) Natural chemistry of Caspian Sea waters. Hdb Env Chem 55:83–108, Springer-Verlag

Wang G, Cheng G (2000) The characteristics of water resources and the changes of the hydrological process and environment in the arid zone of northwestern China. Environ Geol 39:783–790

Wei K, Gasse F (1999) Oxygen isotopes in lacustrine carbonates of West China revisited: implication for post glacial changes in summer monsoon circulation. Quat Sci Rev 18:1315–1334

Wen X, Wu Y, Su J, Zhang Y, Liu F (2005) Hydrochemical characteristics and salinity of groundwater in the Ejina Basin, Northwestern China. Environ Geol 48:665–675

Williams WD, Kokkinn MJ (1988) The biogeographical affinities of the fauna in episodically filled salt lakes: a study of Lake Eyre South, Australia. Hydrobiology 158:227–236

Yang B, Shi Y, Braeuning A, Wang J (2004) Evidence for a warm-humid climate in arid northwestern China during 40–30 ka BP. Quat Sci Rev 23:2537–2548

Zheng X (1987) Salt lakes and their origins in Xinjiang, China. Chin J Oceanol Limnol 5:172–185

Zhu GF, Li ZZ, Su YH, Ma JZ, Zhang YY (2007) Hydrogeochemical and isotope evidence of groundwater evolution and recharge in Minqin Basin, Northwest China. J Hydrol 333:239–251

Zhu GF, Su YH, Feng Q (2008) The hydrochemical characteristics and evolution of groundwater and surface water in the Heihe River Basin, northwest China. Hydrogeol J 16:167–182

Zubakov VA (2001) History and causes of variations in the Caspian Sea level: the Miopliocene, 7.1–1.95 million years ago. Water Resour 28:249–256