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A two-step expansion of the dinocyst Lingulodinium machaerophorum in the Caspian Sea: the role of changing environment

S.A.G. Leroy, H.A.K. Lahijani, J.-L. Reyss, F. Chalié, S. Haghani, M. Shah-Hosseini, S. Shahkarami, A. Tudryn, K. Arpe, P. Habibi, H.S. Nasrollahzadeh, A. Makhlough

Abstract

Dinoflagellate cyst assemblages were analysed in four short sediment cores collected in the south Basin of the Caspian Sea for assessing environmental changes over the last few millennia. Two of these cores were dated by radionuclides. The sedimentation rate of one of them was very high, in the order of 20 mm per year. The interpretation of the four sequences is supported by a collection of 27 lagoonal or marine surface sediment samples. A sharp increase in the concentration of the dinocyst occurs after 1967, especially owing to Lingulodinium machaerophorum. Considering nine other cores covering parts or the whole of Holocene, it became clear that this species started to develop in the Caspian Sea only during the last three millennia. By analysing instrumental data and collating existing reconstructions of sea level changes over the last decades, we show that the main forcing for the recent increase of L. machaerophorum percentages and concentration is global climate change, especially sea surface temperature increase. Sea level fluctuations likely only have a minor impact. We argue that the recent increase in L. machaerophorum indicates that the Caspian Sea clearly is in the Anthropocene.

1. Introduction

The Caspian Sea (CS), with a surface area larger than that of the British Isles, is the largest inland water body in the world (Fig. 1). This lake is well known for the petroleum and caviar it produces. Its ecosystem, which is unique owing to many endemic species, is not available. The instrumental records of the last century indicate rapid changes in sea surface temperature, in sea level and in trophic levels. Sea level changes are large, close to 1 m amplitude and a hundred times faster than that of the global ocean in the 20th century (Crétaux and Birkett, 2006; Fig. 2A), but also as much as 20 m amplitude in the last few millennia (Kakroodi et al., 2012). The sea surface temperature of the south basin especially in the summer shows a warming trend over the last century (Ginzburg et al., 2005).

Using four short sediment cores (36–166 cm long) and 27 surface samples taken in the south basin and adjacent areas, the aims of this paper are 1) to establish a preliminary spatial distribution...
pattern of dinoflagellate cysts in the Caspian area, especially the south basin; 2) to detect and explain changes over the last century in dinocyst assemblages and 3) set it in the context of the Holocene – Neocaspian period. Prime focus will be on the stepwise changes in percentages and concentration of the autotrophic dinocyst *Lingulodinium machaerophorum* (Deflandre and Cookson, 1955) Wall, 1967, subsequently called *Lmac* (abbreviation as in Zonneveld et al., 2013), which is a euryhaline coastal planktonic species restricted to regions with summer temperatures above 10–12 °C and winter temperatures above 0 °C (Lewis and Hallett, 1997). Its motile form, *Lingulodinium polyedrum* (Stein 1883) Dodge 1989, is reported to cause harmful algal blooms (Howard et al., 2009). This species therefore is sensitive to sea surface temperature and could be used to reflect climatic change in the region.

The impact of this work is important in advancing knowledge of CS processes. The CS holds a dominant position in the southwestern Asian region, and is suspected of playing a major part in climate change in this part of the world. However, a lack of a clear understanding of the processes involved in its control such as the North Atlantic Oscillation or the El-Niño Southern Oscillation still exists not only for the last few millennia but also for the last few decades (Arpe et al., 2000, 2012; Arpe and Leroy, 2007). This paper addresses the issue of past temperatures and develops basic data that will underpin future work on CS climate-related aspects.

2. Setting

For two out of the four coring locations more local information is provided in Appendix A.

2.1. Geographical setting and climate

The CS is the world’s largest inland water body in terms of both area and volume, extending 35–48° N and 47–55° E. The altitude of the surface lays around 26 m bsl (25–29 m bsl during the last 150 yr) (Leroy et al., 2006; Arpe et al., 2012). The sea is divided into three basins, becoming deeper southwards: the northern basin (80,000 km²) with an average depth of 5 m and a maximum depth of 175 m; the middle basin (138,000 km²) with an average depth of 175 m and a maximum depth of 788 m; and the southern basin (168,000 km²) with an average depth of 325 m and a maximum depth of 1025 m. The southern basin holds more than 65% of the Caspian water. In 1951, the Caspian Sea was artificially reconnected to the Black Sea via the Volga-Don canal open.

Because of its great meridional extension (>1100 km), the CS straddles several climatic zones (Kosarev, 2005). The northern part of the drainage basin lies in a zone of temperate continental climate with the Volga catchment well into the humid mid-latitudes. The western coast features a moderately warm climate, while the southwestern and the southern regions fall into a subtropical humid climatic zone. The eastern coast has a desert climate.

2.2. Currents, salinity and temperature

A complex sea current pattern in the middle and south basins is dominated by a surface cyclonic gyre (Zenkevitch, 1963). Specifically in the South Caspian, a dipole system, consisting of an anticyclonic gyre in its northwestern part and a cyclonic gyre in its southeastern part, exists throughout the year (Zaker et al., 2011). The sea currents influence the four studied sites as they have potential to transport fine-grained materials. In contrast to the east coast of the middle basin, no upwelling develops in the south basin (Tuzhilkin and Kosarev, 2005).

In general, at the scale of the CS, the summer gradient of salinity is stronger than that of the winter. The main influences are, on an otherwise relatively stable salinity around 11–12 psu, the freshening influence of the Volga River in the north basin (salinities less than 1 psu), and the stronger evaporation in the southeast (salinities up to 14 psu) (Kosarev and Yablonskaya, 1994; Dumont, 1998). Monitoring from 1956 to 2000 shows high salinities in the early 1970s and low salinities in the early 1990s, corresponding respectively to low annual river discharge and high annual river discharge as well as the 1977 lowstand and the 1995 highstand (Fig. 2A). This water level change caused respectively a mixing of the water column and a stratification of the water column (Tuzhilkin and Kosarev, 2005).

In the south basin, the summer and winter gradients show that the west is fresher (12.2 psu) and the east more saline (13.8 psu).
In the CS, summer temperature varies from 26 °C in the NW to >28 °C in the SE. Winter temperatures are 10 °C in the centre of the basin and get cooler by 2–3 °C towards the periphery. A record of sea surface temperature in the south CS from 1982 to 2011 obtained by satellite (Fig. 2B) shows a clear warming trend for all seasons (0.10 °C yr⁻¹) by 1–2 °C and stronger for the south basin of the Caspian than the other regions (Ginzburg et al., 2005). These authors suggest that this warming trend could be a consequence of global warming (Fig. 2C). The sea ice season and extent in the northern basin have decreased for the period 1978–2002, also possibly related to global warming (Kouraev et al., 2004).

2.3. Nutrients and phytoplankton

Most nutrients enter the CS via the Volga River, hence the northern basin has the highest primary productivity. In the CS, the relatively low nutrient levels (Dumont, 1998) in the upper 100 m

**Fig. 2.** Selected oceanographic and meteorological data for the Caspian Sea. All data are smoothed with a 1–2–1 filter. Anomalies are used for panels 2B, 2C and 2D. (A): Caspian Sea level (CSL) (Golitsyn and Panin, 1989; Cazenave et al., 1997; USDA, no date) and summer surface salinity from the centre of the middle basin (sali) (Tuzhilkin and Kozarev, 2005). (B): Sea surface temperatures of southern Caspian Sea from analysis (TCSy = CS year-Jan.–Dec.; TCSs = CS summer-JJA) (Dee et al., 2011); Annual mean 2 m temperatures of the northern hemisphere (Tnh) and annual mean of global ocean temperatures (Toc) (NASA, no date). (C): 2 m temperatures in summer (JJA) for the following meteorological stations (Climatic Research Unit, University of East Anglia, no date): ANZA = 407180, 37.5N 49.5E, –26 m asl, Anzali, Iran LANK = 379850, 38.7N 48.8E, –12 m asl, Lankaran, Azerbaijan DERB = 374700, 42.0N 48.3E, –19 m asl, Derbent, Russia TURK = 385070, 40.1N 53.0E, 82 m asl, Turkmenbashi, Turkmenistan. (D): Annual 2 m temperatures for stations in 2C (Climatic Research Unit, University of East Anglia, no date).
are depleted by phytoplankton activity, but the nutrient concentrations increase with depth (Kosarev and Yablonskaya, 1994). Iranian lagoons and coastal regions have moreover been steadily polluted by anthropogenic sources (fertilizers and pesticides used in agriculture and increased nutrient load of river flows due to deforestation of woodland) since the early 1980s (Kideys et al., 2008).

Changes in phytoplankton biomass are one way to assess changes in water quality. Kideys et al. (2008) detected by satellite imagery an increase in chlorophyll a in the southern CS since 2001. The invasion of the comb jelly Mnemiopsis leidyi in the late 1990s caused a drop in zooplankton, which in turn favoured phytoplankton. But other factors such as overfishing, eutrophication and climatic change may also have played a role (Kideys et al., 2008). A similar analysis for the whole of the CS (Kavak, 2012) shows clearly a positive trend in chlorophyll a from 1998 to 2009.

The status of the south coast of the CS has changed between 1994 and 2005 from oligotrophic to meso-eutrophic (Nassrollazadeh et al., 2008a,b; Bagheri et al., 2012). In 2006, an anomalous algal bloom, mostly due to a dinoflagellate Heterocapsa, occurred in front of Anzali harbour (Bagheri et al., 2011). Other algal blooms mostly due to the toxic Cyanobacteria Nodularia spumigena occurred in August and September 2005 (20,000 km² of the southern basin) (Soloviev, 2005), in August 2009 (off Tonekabon) and in early and early August 2010 (from Nowshahr to near Babolsar and offshore Anzali) (Makloughi et al., 2011; Nassrollazadeh et al., 2011). It is noteworthy that L. polyedrum contributed significantly to the 2009 bloom (Nassrollazadeh et al., 2011).

Despite these environmental changes, the CS is moderately meso-eutrophic, and mainly since 2001. This increase is mostly limited to its coastline, closest to densely populated areas.

2.4. Sediment sources and sedimentation rates

At present, rivers flowing to the northern Caspian coast (the Volga, Ural, Terek and Sulak) supply approximately 90% of the riverine water influx to the CS, whereas the southwestern and southern rivers (the Sefidrud, Kura and Gorganrud) are the main sources of sediments for the whole of the CS (Lahijani et al., 2008).

Studies of the CS sediment composition have so far focused on the northern basin, where the proportions of the different inputs were estimated to be 35% fluvial, 28% aeolian, 29% biogenic carbonate and 8% chemical carbonates (Khristalyov and Artiukhin, 1992). The Iranian mountains, such as the Elburz Mountain, represent the main source of terrigenous materials in the south basin. The sediment has two main primary detrital sources, more siliciclastic in the west, more carbonates in the east (Lahijani and Tavakoli, 2012). One other source is aeolian transport by numerous dust storms from Turkmenistan, where Mesozoic limestones exist in the Kopet Dag (Lahijani and Tavakoli, 2012).

The south part of the basin was subsiding at a rate of the order of > 1 mm yr⁻¹ (Einsele and Hinderer, 1997) and this is probably still the case today.

The sedimentation rates in the south basin have always been very high due to its large accommodation space, strong erosion of the surroundings and orogenesis as well as aeolian inputs. For the Cainozoic, the sedimentary deposits in the southern Caspian basin are as much as 20–30 km thick, making it one of the deepest basins in the world. For the Pliocene–Quaternary time alone, 10 km of sediment have been deposited, providing an average accumulation rate close to 2 mm per year (Nadrov et al., 1997; Tagiyev et al., 1997). The sedimentation rate of the Pliocene in the south basin (the Pliocene Productive Series) is even higher up to 4 mm yr⁻¹ (Kroonenberg et al., 2005).

2.5. Past dinocyst investigations in the Caspian Sea

Some modern samples from the Caspian region were published in Marret et al. (2004), which is also the reference paper for the description of a new Caspian genus, species and forms. Five surface samples were studied for their dinocyst content in the lagoon of Anzali (Kazanci et al., 2004) (Fig. 1).

The recent dinocyst history of the area is poorly known from previous dinocyst investigations, although four records cover the last 300 years at least (locations in Fig. 1): in the Amirkola and Anzali Lagoons (Leroy et al., 2011), in the Gomishan lagoon, S–E corner of the CS (Leroy et al., 2013) and in the Karabogaz Gol (Leroy et al., 2006).

3. Materials and methods

3.1. Collection of surface samples, coring and dating

Modern samples were derived from a combination of core tops, grabs and hand scooping (Appendix B). Sixteen samples came from marine settings, two from lakes, and nine from lagoons (locations in Fig. 1). Two Usnels boxes (50 cm thick) were taken in the south basin, during a French-Russian oceanographic cruise (August 1994), on board a Russian military ship, rented for the sea cruise (Appendix C). Box locations were in water depths of 13 m for core US01 and of 315 m for core US02. Cores from Usnel boxes were subsampled in PVC tubes on board and therefore contain the water—sediment interface. Two heavy Kajak gravity cores with a diameter of 5 cm, cores CS03 and CS10, were taken in 2007 from a boat rented by the Iranian National Institute for Oceanography (INO) (Appendix C). Two of the four cores were dated by radionuclides, i.e. US02 and CS03. ²¹⁰Pb, ²²⁶Ra and ¹³⁷Cs records were obtained. Samples of core US02 were analysed every cm for both ²¹⁰Pb and ¹³⁷Cs in the top 10 cm, then every 2 cm down to 16 cm (results in Leroy et al., 2007). A sample at 22 cm reaches background values. For core CS03, samples were taken at c. 3 cm interval down to 60 cm and then at every c. 8 cm interval. Dry samples were measured with a very low-background, high-efficiency well-type detector located in the underground laboratory of Modane in the French Alps (Reyss et al., 1995) where 1700 m of rock overburden reduces the cosmic radiation by 6 orders of magnitude.

3.2. Sedimentology

For the sedimentary analysis of the two Usnels cores, dried samples were homogenized and representative subsamples were taken for grain size analysis by a LS 13 320 Multi-Wavelength Particle Size Analyzer, ASTM standard calibrated. Organic matter and calcium carbonate contents were determined by loss-on-ignition, bulk samples were dried at 105 °C for 24 h, then heated to 550 °C for 4 h to burn the organic matter, and organic free sample heated again to 950 °C for 8 h to break the calcium carbonate. The percentages of organic matter and calcium carbonate were then calculated by the method used by Dean (1974).

For the two Kajak cores, samples were homogenized and representative subsamples were taken for grain size analysis. The distribution for the fraction coarser than 1 mm was determined using the standard wet sieving procedure. Grain-size analysis for particles less than 1 mm was undertaken using a “Fritsch Analysette Comfort 22” Laser Particle Sizer. Organic matter was determined by wet digestion through oxidation in hydrogen peroxide on bulk
samples (Schumacher, 2002). The calcium carbonate was determined by using a Bernard calcmetre.

The magnetic susceptibility was measured with a Bartington Instruments MS2 susceptibility bridge. Cores CS03 and CS10 were measured using a MS2C sensor on half cores, core US01 using MS2E1 and core US02, using MS2F sensor, both along flat surface of half cores. The thermomagnetic behaviour of the bulk sediment sample from core US02, was determined on a horizontal force translation balance in air atmosphere in a magnetic field of 0.375 T and with a linear temperature increase.

### 3.3. Palynology

Samples were taken every 5 cm in the four cores. Initial processing of samples (1–2 ml) involved the addition of cold sodium pyrophosphate to deflocculate the sediment. Samples were then treated with cold hydrochloric acid (32%) and cold hydrofluoric acid (60%), followed by a further treatment with hydrochloric acid. The residual organic fraction was then screened through 125 and 10 μm mesh sieves and mounted on slides in glycerol. Lycopodium tablets were added at the beginning of the process for concentration estimates, which are provided in number of dinocysts per ml of wet sediment (c ml⁻¹).

The dinocysts were counted alongside pollen and other microfossils. All palynomorphs appeared well preserved. The sum for percentages is made of all dinocysts. Varia are expressed as a percentage of the same sum as the other dinocysts. The percentages of the foraminiferal inner organic linings found in the palynological slides were displayed in the same way. A ratio pollen concentration on dinocyst concentration (P:D) has been determined according to McCarthy and Mudie (1998) in order to estimate the degree of continentality of the assemblage.

A statistical analysis available in Psimpoll (Bennett, 2007), the zonation by cluster analysis (CONISS) after square root transformation of the percentages, was applied. The zonation was calculated for the percentage diagrams. The taxonomy of the few Caspian dinocyst species has been established by Marret et al. (2004) including three endemic species. Illustrations of Caspian dinocysts are also available in Leroy (2010) and Mudie et al. (2011). Three forms of Lmac have been defined: ss, A and B, in the KaraBogaz Gol; this is also used for the all the Lmac in the CS (Leroy et al., 2006). Form A has numerous, shorter (average 9 mm in length) and slender acuminate processes than A and B, in the KaraBogaz Gol; this is also used for the all the other forms, from at least 8.5 to 42 psu, and for temperature from 0 to 31 °C, and with a linear temperature increase.

### 4. Results

#### 4.1. The surface samples

The samples are organised by increasing percentage values of Lmac B from the top to the bottom of Fig. 3, except the bottom two samples which have high values of forms ss and A.

At the top of the diagram, the assemblages of the two lake samples (Amaqol and Alagol) are composed only of Brugantedinium spp. At the bottom of the diagram, the dinocysts are dominated by Lmac B with increasing values of Lmac ss and often the presence of Spiniferites belerius (saline lagoons of BTork 1 and 2, and TR1). The bottom-most two samples are derived from hypersaline settings: the Aral Sea (AS17-5) and the Karabogaz Gol (KBG8-01) explaining respectively the abundance of Lmac ss and the high peak of Lmac A.

In the middle sample group, the spectra are dominated by Impagidinium caspienense and Caspindium rugosum rugosum, with occasionally at the top end of the diagram the development of cysts of Pentapharsodinium dalei and Spiniferites cruciformis, respectively in the middle Caspian basin (cores US24 and US26) or more freshwater lagoons (Anzali).

#### 4.2. Core sediment and dating

The sediment of core US01, off the coast of Turkmenistan, is silty with up to 20% clays. It has the highest amount of sand of the four cores, reaching 10–20% (Fig. 4). The core bottom has 4% organic matter, but this rapidly decreases upwards to 1%. The carbonate content is stable around 20%. The magnetic susceptibility is very low and increases upwards. Benthic foraminiferal tests are abundant (Appendix D).

The sediment of core US02, from the middle of the south basin, is clayey and silty with some sand from 8.5% at the base to 2% at the top. The organic matter progressively increases from 2 to 4%. This sequence has the highest content in carbonates of the four cores, reaching 30–40%. It is likely that some of the carbonate is derived from benthic ostracod valves as they are known in the top cm of this core (Boomer et al., 2005). The magnetic susceptibility is low. The Curie temperature for sample at 6 cm depth was 580 °C; the mineral carrying the magnetic susceptibility is magnetite (Fe₃O₄).

The sediment of core CS03, offshore from Anzali, is comprised of grey silts and clays with darker layers. The sediment is occasionally bioturbated by a network of animal tubes. The mean amounts of organic matter and carbonates are stable with respectively values of 3 and 17%. The magnetic susceptibility is low.

The sediment of core CS10, taken offshore from Babolsar, consists of clayey silts, sometimes with very fine sand, poorly sorted, getting darker with depth, ranging from dark grey to black with a reduced-sediment odour. Sediments are laminated through much of the sequence. Some part of the sequence appears burrowed (particularly the lower portion). The sequence has 3% of organic matter and 20% of carbonates as well as the highest magnetic susceptibility of the four cores.

The radionuclide results of core US02 have been published in Leroy et al. (2007). In brief the sedimentation rate obtained is 2.0 mm yr⁻¹ for the top 22 cm of sediment. The base of the core is therefore estimated to be at AD 1709.

For core CS03, the record of artificial fallout radionuclides ²⁴¹Am and ¹³⁷Cs (Fig. 5) detected in the upper 104 cm of sediment displays a maximum at 83 cm depth that is assigned to the period of maximum weapons fallout, i.e. AD 1963 (Appleby, 2000). The corresponding averaged sedimentation rate is thus of 19 mm yr⁻¹. The radiometric ²¹⁰Pbex profile (Fig. 5) does not display the regular exponential decrease with depth in the core as expected for a constant
Dinocyst assemblages from marine and lagoonal surface samples, mostly from the south basin of the Caspian Sea, topmost samples from two lakes and bottom-most samples from high salinity environments. Percentages are on the sum of all dinocysts. For the foraminiferal linings the sum includes all dinocysts except the foraminifera. Concentration in numbers of dinocysts per ml of wet sediment.

Fig. 3. Dinocyst assemblages from marine and lagoonal surface samples from high salinity environments. Percentages are on the sum of all dinocysts. For the foraminiferal linings the sum includes all dinocysts except the foraminifera. Concentration in numbers of dinocysts per ml of wet sediment.

4.3. Dinocysts in the four cores

In general the same range of taxa is found in the four cores (Fig. 6), with the usual dominance of I. caspienense and the abundance of Lmac under two forms: B and ss (Marret et al., 2004). Pyxidinopsis psilata has only rare occurrences.

4.3.1. Core US02, the last 300 years in the middle of the south basin

Zone US02-d1 has relatively high S. cruciformis values and a significant percentage of C. rugosum, not met in the surface samples. Zone US02-d2 has slightly less S. cruciformis, and after 10 cm much less C. rugosum and slightly more Lmac B.

The concentration is the lowest of the four cores due to the distance to the shores. The relatively low P:D displays an increasing overall trend. This diagram shows a mild trend to have opposite fluctuations of P. dalei cysts (highest values of the 4 cores) and Lmac B. This core overall shows assemblages that indicate the lowest salinities of the four cores. This is due to its location in a more open setting. No foraminiferal linings have been found in significant numbers throughout the core and form a continuous curve.

4.3.2. Core US02, the last 300 years in the middle of the south basin

Zone US02-d1 has relatively high S. cruciformis values and a significant percentage of C. rugosum, not met in the surface samples. Zone US02-d2 has slightly less S. cruciformis, and after 10 cm much less C. rugosum and slightly more Lmac B.

The concentration is the lowest of the four cores due to the distance to the shores. The relatively low P:D displays an increasing overall trend. This diagram shows a mild trend to have opposite fluctuations of P. dalei cysts (highest values of the 4 cores) and Lmac B. This core overall shows assemblages that indicate the lowest salinities of the four cores. This is due to its location in a more open setting. No foraminiferal linings have been found in this sequence as the water depth is too great for the survival of these benthic organisms in the CS (Boomer et al., 2005).

4.3.3. Core CS03, the 20th century off Anzali

In zone CS03-d1, I. caspienense has maximal values, P. dalei cysts are well represented and the percentages of Brigantedinium are relatively high but decreasing. The concentration increases slowly across this zone while the P:D decreases upwards. In zone CS03-d2, Lmac B values are nearly twice as high as before. P. dalei cysts become rare. The concentration is clearly higher than in zone 1, while P:D is stable. This core shows a compelling opposition between P. dalei cysts and Lmac B.

The coring location, 15 km offshore, does not seem to be influenced by the freshwater outflow from the Anzali lagoon.
Fig. 4. Sedimentology of the four short cores: visual description of the lithology (dotted lines for sand, dash-dot lines for silt, horizontal lines for thin black layers, wiggles for bioturbation, and grey comas for shells), organic matter (OM) and calcium carbonate (CaCO₃) contents in percentages after Loss-On-Ignition, clay, silt and sand percentages by particle size analysis, as well as magnetic susceptibility (MS).

Fig. 5. Radionuclide dating of core CS03. (A): Profiles of ²¹⁰Pb activities (Bq kg⁻¹, dry weight) with depth in cm. The regression line between the top and 65 cm depth corresponds to an average sedimentation rate of 2.0 cm yr⁻¹. Deeper in the core the regression line corresponds to 3.8 cm yr⁻¹. (B): Profiles of artificial radionuclides ¹³⁷Cs and ²⁴¹Am activities (fallout from nuclear weapon testing) with depth in cm. The maximum activity at 83 cm depth for both nuclides corresponds to the maximum global fallout peak in 1963. Vertical bars for error.
Caspian South basin analysis: S. Leroy

off Turkmenistan, core US01

off Anzali, core CS03

off Babolsar, core CS10

Fig. 6. Dinocyst diagrams of cores US01, US02, CS03 and CS10, taken in the south basin of the Caspian Sea. Concentration in numbers of dinocysts per ml of wet sediment. With 10x exaggeration curve. To the right dendograms used for the zonation.
4.3.4. Core CS10, with massive dinocyst concentration increase off Babolsar

In zone CS10-d1, the values of I. caspienense are the highest of the four cores. The P:D is extremely high, showing a clear terrestrial influence similar to that in surface samples, e.g. in lakes (such as Alagol) and lagoons (Anzali and Bandar-e-Torkman).

In zone CS10-d2, after a sudden change, high percentages of Lmac B (80%) occur. These are also the highest values ever recorded in the CS. This occurs in parallel to a huge increase in the dinoflagellate cyst concentration.

The limit between zones CS03-d1 and d2 corresponds well to that between zones CS10-d1 and d2, with the same increase in gellate cyst concentration. In the CS, the Roman number II indicates the recent expansion. Core KBG08 (Leroy et al., 2006); cores GS05 and GS18 (Leroy, pers. comm.); Core TM (Leroy et al., 2013; the arrow indicates that only the part with Lmac in the 27.5 m long core is represented); cores HCGL02 and HCGA05 (Leroy et al., 2011); cores CP14, CP18 and CP21 (Leroy et al., 2007). Dates in AD or in calibrated years BP.

4.4. Concentration of Lingulodinium machaerophorum in thirteen Caspian Sea cores

In order to put in context the changes in Lmac observed in the four short cores, the records of Lmac concentrations in thirteen cores analysed by the same palynologist across the south and middle basins of the CS are compared (Fig. 7).

It is clear that the lagoons yielded the highest concentrations in Lmac often higher than 5000 cysts per ml of wet sediment (c ml⁻¹). Occasionally extraordinary values are even reached, such as at the top of the coastal core CS10 (>20,000 c ml⁻¹), at the base of the Anzali lagoon (HCGA05) and in the KaraBogaz-Gol (up to 141,000 c ml⁻¹). On the contrary, the cores taken in the middle of the basins display low values, even lower than 1000 c ml⁻¹, such as CP14 and US02 in the south basin and CP21 in the middle basin.

5. Discussion

5.1. Dinocyst distributions

The spectra from the surface sediment and the four cores are largely dominated by I. caspienense and Lmac. The near absence of P. psilata and the low values of S. cruciformis (for the latter, except the central core and some surface samples in the Lagoon of Anzali) are a characteristic of these four recent records and of the surface samples. P. psilata is mostly found in Khvalynian (Late Pleistocene) sediment (S. Leroy, unpublished data) and in lagoons and river deltas such as the Demchik area of the Lower Volga River (K. Richards, pers. comm.). This is easily explained both in terms of the higher salinity of the Neocaspian Sea and in terms of survival in lagoons where the salinity is variable and where the lagoons maintain at most times small areas of lower salinity if fed by rivers. The higher values of S. cruciformis in zone US02-d1 may reflect the much higher water levels of the LIA, which also have less saline waters as also seen in the lagoon of Anzali (Leroy et al., 2011).

The two southern cores have the highest percentages of Brigantedinium. This is explained by the proximity of a densely inhabited coastal area producing a lot of nutrients discharged into the sea. The cosmopolitan species P. dalei cyst is clearly more abundant in the northern surface samples (samples US24 and 26), which are characterised by larger amplitudes of sea surface temperatures and the proximity of sea ice in winter (Marret et al., 2004).

On the one hand, based on the locations of the cores and the surface samples, it is possible to show that the three Lmac forms...
show an increasing salinity gradient from B, to ss and finally to A. On the other hand, *Lmac* percentages (all forms included) display an increase across the CS to the southeast, i.e. towards regions that are warmer and more saline. So the morphology of the cysts responds to the salinity (shown by another method in Mertens et al., 2009) but its biomass to temperature.

5.2. Exceptional sedimentation rate

The sedimentation rate of core US02, in the centre of the south basin, obtained by the radionuclide method is 2 mm yr⁻¹. This is 10 times higher than that obtained by radiocarbon on core CP14 taken close by Leroy et al. (2007). This difference could be caused not only by an expected lack of compaction of the top tens of cm of sediment in core US02 but also by a very recent (the last few centuries) increase of the sedimentation influx (Leroy et al., 2007). This radiocarbon-based value falls within the sedimentation rates of other cores from the deep basins, i.e. between 0.02 and 0.54 mm yr⁻¹ (Amini et al., 2012).

The sedimentation rate in a short core (HCGA05, 170 cm long) from the Anzali Lagoon, is of 5 mm per year, that of core HCGL02 (95 cm long) in the lagoon of Amirkola is 2.5 mm yr⁻¹ (Leroy et al., 2011), and that in cores from the Gorgan Bay is between 1.4 and 2.45 mm per yr (Karbassi and Amirmeshid, 2004). These values are higher than those from the core from the centre of the south basin. These lagoonal settings are indeed expected to have higher sedimentation rates than those of the open sea.

On the whole the values of the short cores are in the order of the sedimentation for the Cainozoic (Nadirov et al., 1997; Tagiyev et al., 1997).

However 20 mm per yr obtained on the coastal core CS03 is truly exceptional (Amini et al., 2012). This value is even higher than that of the Pliocene Productive Series (Kroonenberg et al., 2005). This is explained by the core location on the slope of the continental shelf. The time resolution of the palynological samples therefore reaches one sample every 2.5 years.

5.3. Shifts in *L. machaerophorum* concentration

This section focuses on what are the possible causes for the increase in *Lmac* visible in the four cores, more so for the two southern ones (Fig. 6 and marker II in Fig. 7). Moreover these increases of *Lmac* are often to the detriment of *P. dalei* cysts when present in the cores. The recent changes are then set in the context of the development of *Lmac* in the CS in the last few millennia (marker I in Fig. 7).

5.3.1. In the last decades

*L. polyedrum*, the motile form of *Lmac*, usually occurs in the water column during late summer, which suggests that a minimum temperature is needed for its development. *L. polyedrum* can bloom in nutrient rich and nutrient depleted waters: thus its distribution does not have to be restricted to areas with high nutrient concentrations in surface waters (Lewis and Hallett, 1997). Dinoflagellate biomass that shows a steep increase as part of anomalous algal blooms that occurred in the last few years in the CS are probably seen in the extreme dinocyst concentration increase across core CS10, especially since 1967 (marker II in Fig. 7). The causes for this are multiple, and are often attributed to nutrient washed into the sea by the rivers; however they may also be specifically linked to higher temperature and low wind conditions and therefore stratified waters (Soloviev, 2005; Nasrollahzadeh et al., 2011). A remote sensing analysis for the 2005 large-scale phytoplankton bloom indicated that an increase of 4 °C was observed in the bloom itself in comparison to surrounding waters. Low wind conditions and water stratification were also observed. In brief this would tend to indicate that *L. polyedrum* responds primarily to high temperatures, with salinity and nutrients as secondary factors.

In the 20th century, the water level has fluctuated by ±3 m a couple of times (Fig. 2A), but this seems to have been hardly registered in the fossil record (Fig. 6). More decisively, the sharp sea level rise between 1977 and 1995 has led to a salinity decrease. However *Lmac* percentages have strongly increased. Therefore this is more in line with the gradual increase of sea surface and air temperatures throughout this period (Fig. 2B–D).

5.3.2. In the Late Holocene

Successive dinocyst phases are characterised by the dominance of various dinocysts in the later part of the Late Pleistocene—Holocene history of the CS: 1) the most recent is a *Lmac* phase in the last few decades, 2) the next one is dominated by *l. caspisen* until c. 3.9 cal. ka ago, and 3) the oldest one, covering the early Holocene and Late Pleistocene, is dominated by *S. cruciformis* and *P. psilata*. This succession was already suspected in core CP14 (Leroy et al., 2007), but it is only with the additional cores US01, US02, CS03 and CS10 that the most recent phase became clearly visible.

The examination of thirteen records reveals the following history in the development of *Lmac* (Figs. 1 and 7). It is quasi absent from a 10 m long core in the centre of the south basin, core GS05, whose top reaches c. 4000 yr ago (Pierrret et al., 2012; S. Leroy, unpublished data) and only prevails in the top cm of a 40 m long core in the centre of the middle basin, core CS18, from c. 2900 yr onwards (Boomer et al., 2005; S. Leroy, unpublished data) (marker I in Fig. 7). In the Holocene lagoonal core TM in Gomishan, SE of the south basin, *Lmac* starts to develop only from c. 3200 yr ago (Leroy et al., 2013). In the marine cores CP14, 18 and 21 covering most of the second half of the Holocene (Leroy et al., 2007), a steady increase is observed. More especially in core CP14, *Lmac* starts to develop c. 2500 yr ago (Leroy et al., 2007) (marker I in Fig. 7). The other much shorter records, KBG08, HCGL02 and HCGA05, show the abundance of this dinocyst in the last few centuries (Leroy et al., 2006, 2011). In core HCGA05 of the Anzali Lagoon (Leroy et al., 2011), *Lmac* fluctuates in anti-phase with *S. cruciformis*; this was interpreted as a salinity signal. The highest percentages of *Lmac* B, 80%, and *Lmac* A, 80%, are respectively found in core CS03 (Iranian coast) and in the Karabogaz-Gol.

Contrary to the CS, in the Black Sea, *Lmac* is already present at the beginning of the Holocene. In the SW of the Black Sea, *Lmac* occurs in cores MO2–45 and MAR05–13 respectively over the last 9500 uncal. ¹⁴C years and 11,000 uncal. ¹³C years (Marret et al., 2009; Bradley et al., 2012) (note that the Black Sea radiocarbon dates are often uncalibrated and uncorrected for the reservoir effect due to major uncertainties). From 6 uncal. ¹⁴C ka, it is even largely dominant. It peaks from 6 to 5 uncal. ¹⁴C ka, which is considered the warmest period in the climate of the Black Sea region. The topmost part of the core estimated to be close to the present shows at least 80% of *Lmac* ss. Marret et al. (2009) suggested that these high values were due to anomalous blooms of this species.

In a Black Sea sedimentary sequence, Verleye et al. (2009) found an opposition, similar to that of the present study, between *Lmac* and *P. dalei* cysts interpreted here as less river input versus higher terrigenous input. In the Sea of Barents, the abundance of cysts of *P. dalei* increases according to Voronina et al. (2001) near high productivity waters. An investigation of dinocysts in a core from the Marmara Sea covering the Late Pleistocene and Holocene considers *Lmac* as a thermophilous taxon (Londeix et al., 2009).

*Lmac* is present with high percentages shifts in the 2000 yr-long record obtained from the Aral Sea (Sorrel et al., 2006). Although the authors suggest that the salinity is the first forcing factor, they recognise that the temperature in the Tian Shan Mountains (by control on snow melt) are the ultimate driver.

In brief in the CS, contrary to the Black Sea, it appears that the *Lmac* development started slowly only c. 3200 yr ago (marker I in Fig. 7) and has strongly accelerated in the last decades (marker II in Fig. 7).
5.3.3. Possible causes

The causes for the development of \( Lmac \) in the CS c. 3200 yr ago could be linked to the instauration of favourable regional conditions, a possible long-term recovery from glacial meltwater inflow (Leroy et al., 2007) or at a global level. It is not an appearance as the species was already found in extremely low numbers in the CS in the Holocene (cores GS05 and GS18), as well as in the Late Khazarian in Tsagan-Aman (Tudryn et al., 2012) and in the Pliocene (K. Richards, pers. comm.).

Changes in phytoplankton abundance may be driven by several factors that may confound the climatic signal (Adrian et al., 2009). In the present situation, it is however possible to eliminate two important factors that may have explained the changes in \( Lmac \): sea level changes and eutrophication. The causes of the second and more recent development of \( Lmac \) are not sea levels that have fluctuated widely during that time period. During the LIA, the water level was 6 m higher than at present (palaeoenvironmental, archaeological and documentary evidences; Brückner, 1890; Leroy et al., 2011; Naderi et al., 2013a). During the early Middle Ages, the water levels were lower by 2–4 m as attested by the numerous walls built by the Sasanian Empire (archaeological and documentary evidences in Brückner, 1890; Omrani Rekavandi et al., 2007). A highstand similar to the LIA has been recorded between 2600 and 2300 cal. yr BP (sediment: Kroonenberg et al., 2007; Naderi Beni et al., 2013a,b). These significant changes have so far not been recorded in the \( Lmac \) curve. So sea level changes have taken place in the last millennia and even centuries but did not significantly affect the percentages of \( Lmac \). In addition, because the shift to increased values of \( Lmac \) is older than anthropogenic eutrophication and because here \( Lmac \) is inversely correlated to \( P. \) dalei, eutrophication may be secondary only in explaining the CS trend. Moreover it is only the coasts of the CS that are becoming meso-eutrophic, and especially since 2001 (Dumont, 1998; Nezlin, 2005).

The temperature trend across the later part of the Holocene, due to a long-term recovery from the end of the meltwater inflow into the CS, is the initial forcing factor. The recent global warming trend detected in the CS 20th century instrumental data (air and sea surface), and confirmed for the first decade of the 21st century (Fig. 2B), has fuelled the second increase of \( Lmac \).

In brief (Fig. 8), it has been shown that the increases of \( Lmac \) 1) some 3200 years ago was due to natural global and/or regional climatic forcing working, such as a long-term recovery from meltwater inflow, and 2) in the last decades was due to recent warming of the surface waters related to a recent anthropogenic-induced global forcing. It is therefore hypothesised that the temperature increase is the primary forcing factor, and salinity increase is just a corollary, reinforced by the closed basin setting.

6. Conclusions

This study is the first to provide an analysis of modern and very recent dinocyst assemblages for the CS, with more details for the south basin in some cores with exceptional sedimentation rates.

This investigation of surface samples has provided a baseline for the present state of the CS and surroundings, with the dinocysts reflecting surface water parameters. This may serve as a reference when comparing it with past states and future ones. More specifically for the present, the following was observed: more \( P. \) dalei cysts in the north at proximity to winter sea ice and with larger seasonal contrasts, more \( S. \) cruciformis in freshwater lagoons or in the oligohaline north, and more \( Lmac \) in the warmer south and in higher salinity settings.

New analyses on four short cores in the CS (with one of them with a sedimentation rate of 20 mm per year) have shown a recent expansion of the dinoflagellate cyst \( Lmac \). When these results are placed in the context of thirteen sequences in the CS, a robust signal is obtained: the dinocyst concentration trend is the end result of a natural increase that has started slightly over three millennia ago. The overall trend has recently been amplified by global warming as shown by instrumental data. This warming (and the subsequent stratification increased) has significantly changed the dinocyst assemblage in terms of relative abundances and in terms of concentration; whereas salinity via sea level changes, which are typical driving factors of this species abundance, were shown here in the Caspian Sea not to have had a significant influence.

The dinocyst assemblages and the sea surface temperatures indicate that the CS has moved into the Anthropocene, i.e. the post-industrialisation period (Crutzen, 2002), as many other lakes (Adrian et al., 2009). The CS is a fragile ecosystem exploited by multiple users who each wish to sustain their economical development and who will have to change in order to adapt to the present changes.

The next steps could be 1) to make culture experiments on the effect of temperature on abundance of \( Lmac \) (e.g. Hallett, 1999), and 2) to measure the lengths of the \( Lmac \) processes in the CS as in the Black Sea (Mertens et al., 2009, 2012). Both would lead to the establishment of quantitative reconstructions of environmental parameters such as temperature and salinity in the Caspian Sea over the Holocene and longer time periods.

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Appendices

Appendix A. Modern oceanographic parameters for Anzali and Babolsar

Offshore from Anzali (one of the four main studied locations, S–SW Caspian coast of Iran, Fig. 1) across the four seasons of the year 2006, the temperature and salinity profiles at a station with 50 m water depth indicate a constant temperature down to 30 m followed by a thermocline (26–9 °C). The salinity (9.32 and 12.25) is also constant, with just a drop in surface waters in February (Bagheri et al., 2011).

The temperature structure offshore from Babolsar (close to another one of the four main studied locations, S–SE Caspian coast of Iran, Fig. 1) is characterised by a strong seasonal thermocline, located between 20 and 50 m depth with a 15 °C temperature difference across it in summer. In autumn, the thermocline gradually weakens and, at the end of winter, it disappears before its reformation in the early spring. The salinity is 12 psu with a slight decrease to 11 psu in enhanced river flow to the sea (Zaker et al., 2007, 2011).

Temperature and salinity measurements along the Iranian coast show similar temperatures but with slightly wider seasonal amplitude in the west (Appendix B). The salinity increases eastwards because of the inflow of freshwater in the west via Anzali lagoon and the Sefidrud River.

No sewage control exists in Babolsar: industrial, domestic and agricultural wastes go to the sea via the river. Moreover, in both the flat coastal plains of Anzali-Rasht and Babolsar, a significant source of nutrients is derived from the fertilizers used in the rice fields.

Appendix B

Table B1

Location of surface samples for dinocysts according to increasing values of *L. machaerophorum* B.

| Label | Lat N | Long E | Water depth in m | Location brief description | Sampling date | Type of sample | Surface temperature in °C | Surface salinity | Date of measurements & source | Published palynological sample label |
|-------|-------|--------|-----------------|-----------------------------|---------------|-----------------|--------------------------|----------------|-----------------------------|----------------------------------|
| Almagol | 37 25 53.50 | 54 38 52.18 | 0.6 | Core top in Modern Lagoon, water at zero | Sept. 2010 | Lagoon | 11–28 | 2–3 | Not measured at sampling; Patimar (2008) for 2000–2002 | |
| Alagol | 37 21 59.48 | 54 34 44.33 | 0.6 | Core top in Modern Lagoon, water at ~6 m | Sept. 2010 | Lagoon | 10–27 | 3.5–4.0 | Not measured at sampling; Patimar (2008) for 2000–2002 | |
| Anzali 09 | 37 26 56.6 | 49 22 49.8 | 1.8 | Grab HCGA09, 280 m away | 26 June 2008 | Lagoon | n/a | n/a | Not measured at sampling | |
| Anzali 6 | 37 25 1.94 | 49 25 17.93 | 1–2 | Grab Anzali Lagoon surface | 1995 | Lagoon | 4.5–27.5 | 0.8–3.0 | Not measured at sampling | EnSS15a |
| US24 | 43 19 14 | 49 06 02 | 61 | Core top | Aug. 1994 | Marine | 25.5 | 9.5 | F. Chalié, snapshot | US02b |
| US02 | 39 16 | 51 28 | 315 | Core top, SR9406US14, SR01US9402, at 0.5 cm depth | Aug. 1994 | Marine | 24.7 | 10.9 | F. Chalié, snapshot | |
| Anzali 3 | 37 29 5.62 | 49 19 14.05 | 1–2 | Grab Anzali Lagoon surface | 1995 | Lagoon | 4.5–27.5 | 0.5–4.5 | Not measured at sampling; but across 2000a | EnSS15a |
| Anzali 100 | 37 36 | 49 32 | 100 | Grab during phytoplankton survey | Jan. 2011 | Marine | 9.3–30 | 7–11 | H. S. Nasrollahzadeh, annual | |
| Astara 20W | 38 24 | 49 01 | 20 | Grab during phytoplankton survey | Winter 2010 | Marine | 7.5–33 | 7.5–14 | H. S. Nasrollahzadeh, annual | |
| Torkman 20 | 37 05 | 53 35 | 20 | Grab during phytoplankton survey | Winter 2010 | Marine | n/a–29.5 | 9–10 | H. S. Nasrollahzadeh, annual | |
| US01 | 38 44 10 | 53 11 15 | 13 | Core top, SR0403US09, SR01US9401, at 0.5 cm depth | Aug. 1994 | Marine | 28 | 10.5 | F. Chalié, snapshot | |
| CS10 | 36 48 25.0 | 52 33 02.8 | 250 | Core top, core CS10 at 2 cm | 2007 | Marine | 18 | 12 | Not measured at sampling; but snapshot for a station with 42 m water depth in Nov. 2008 (Janschidi & BinAbuBakar, 2011) | |
| Astara 100 | 38 22 | 49 08 | 100 | Grab during phytoplankton survey | Summer 2010 | Marine | 7.8–32.4 | 7–12 | H. S. Nasrollahzadeh, annual | |
| Astara 205 | 38 24 | 49 01 | 20 | Grab during phytoplankton survey | Summer 2010 | Marine | 7.5–32.6 | 7.5–14 | H. S. Nasrollahzadeh, annual | |
| Babol 100 | 36 49 | 52 39 | 100 | Grab during phytoplankton survey | Jan. 2011 | Marine | 9.6–28 | 9.5–13 | H. S. Nasrollahzadeh, annual | |
| US26 | 43 19 36 | 49 05 58 | 61 | Core top | Aug. 1994 | Marine | 25.5 | 9.5 | F. Chalié, snapshot | |
under a binocular microscope and

250 for 15 h and were wet sieved through mesh sizes of 53, 63, 125, 125 and 250 μm fractions shows a maximum abundance in the 63

m fractions. The maximal abundance of foraminifera is at 7 cm depth (Fig. D1). In the entire core, A. beccarii maximum abundance is at 14 cm depth and minimum abundance at the 34 cm of core, but the abundance of E. littorale is reversed at these depths.

Table B1 (continued)

| Label      | Lat N | Long E | Water depth in m | Location brief description                  | Sampling date | Type of sample | Surface temperature in °C | Surface salinity | Date of measurements & source | Published palynological sample label |
|------------|-------|--------|------------------|---------------------------------------------|---------------|----------------|------------------------------|-----------------|-------------------------------|-------------------------------------|
| CS03       | 37 35 28.3 | 49 34 16.6 | 250 | Core top, core CS03 at 2 cm | 2007 | Marine | 10 | 12 | Not measured at sampling, but for a station nearby in 2008 | Ens6a |
| Anzali 4   | 37 25 18.16 | 49 26 48.36 | 1–2 | Grab Anzali Lagoon surface | 1995 | Lagoon | 4.5–27.5 | 1.0–5.0 | Not measured at sampling; but across 2000 | Ens6a |
| Anzali 20  | 37 30   | 49 29   | 20 | Grab during phytoplankton survey | Jan. 2011 | Marine | 7–32 | 7–10 | Not measured at sampling; but across 2000 | Ens12a |
| Babol 20   | 36 46   | 52 40   | 20 | Grab during phytoplankton survey | Jan. 2011 | Marine | 10–29.4 | 8–11.4 | H. S. Nasrollahzadeh, annual | Ens1a |
| Anzali 1   | 37 24 55.24 | 49 29 9.12 | 1–2 | Grab Anzali Lagoon surface | 1995 | Lagoon | 4.5–27.5 | 0.6–2.0 | Not measured at sampling; but across 2000 | Ens1a |
| BTorkman 2 | 36 53 57.3 | 54 02 46.1 | 0.1 | Artificial pool behind bay, salinity 22, reeds, Salicornia | 20 May 2011 | Lagoon | n/a | 22 | S. Leroy, snapshot | |
| BTorkman 1 | 36 53 49.7 | 54 02 39.4 | 0.1 | Harbour, salinity 17, mud with Salicornia meadow | 20 May 2011 | Lagoon | n/a | 17 | S. Leroy, snapshot | |
| TR1        | 37 03 43.70 | 54 01 59.02 | 0.1 | Core top in Modern Lagoon, water at ~28 m, Gm1short | Sept. 2010 | Lagoon | 9–30 | 20–24 | Spring–Summer in 2007 (Patimar et al., 2009) | |
| AS17       | 46 31 04 | 60 41 55 | 12.5 | Core top, AS17-5 at 0.5 cm | 17 June 1999 | Marine | n/a | 22 | (Boomer et al., 2003; S. A05b) | |
| KBG08-01   | 41 51   | 53 15   | 0.8 | Core top, KBG08-01 at 0.5 cm depth | 1999 | Marine | 35.5 to 33 | 170–250 | KBG08-01b | |

n/a: not available.

A. beccarii is the most abundant taxon in this core and carii dominates this core and is indicative of a shallow-marine environment with sandy bottom (Sgarrella and Moncharmont Zei, 1993). However below 20 cm depth, E. littorale is more abundant and this could be paralleled to the higher organic matter values.

Appendix C. Core numbering and location. Note: Locations and water depths for the two Usnel cores are approximate owing to poor weather conditions and navigation tool performances.

| Full name | General location | Short name | Length in cm | Water depth in m | Date corer |
|-----------|------------------|------------|--------------|------------------|------------|
| On board no SR9403US09 | US01 | 36 | 38°44′10″ | 13 | 1994 Usnel box |
| Museum no SR01US9401 Offshore Turkmenistan | US02 | 57 | 39°16′ | 315 | 1994 Usnel box |
| On board no SR9408US14 | US01 | 57 | 39°16′ | 51°28′ | 2007 Kajak |
| Museum no SB01US9402 Centre of south basin | CS03 | 166 | 37°35′28.3″ | 250 | 2007 Kajak |
| Offshore Anzali CS07-03 | CS03 | 166 | 37°35′28.3″ | 250 | 2007 Kajak |
| Offshore Babolsar CS07-10 | CS10 | 147 | 36°48′25.0″ | 250 | 2007 Kajak |

Appendix D. The analysis of foraminiferal tests in core US01

Materials and methods

Due to the continuous presence of foraminiferal linings in palynological slides, core US01 was selected for foraminiferal test analyses. Seven samples were taken at approximately 5 cm intervals. The samples were dried at 50 °C in an oven and weighed. Subsequently the samples were treated with a 4% solution of H₂O₂ for 15 h and were wet sieved through mesh sizes of 53, 63, 125, 250 μm fractions and then were dried again. Afterwards, all foraminifera were picked and identified from 53, 63, 125 and 250 μm fractions under a binocular microscope and finally percentages were calculated for each species. In the case of samples rich in foraminifera, the samples were split into 1/8 prior to sieving, then picked and the total abundances were then calculated by multiplying.

The taxonomic classification followed the suprageneric taxonomy of Loeblich and Tappan (1987), with some information from Birshstein et al. (1968).

Results

The sediment contains five different euryhaline benthic foraminifera taxa: Ammonia beccarii (47–87%), Elphidium littorale (also known as Elphidium littorale caspicum, Retroelphidium littorale, Retroelphidium caspicum, Elphidium caspicum caspicum and E. gunteri) (10–52%), Elphidium shochniae (0.9–0.83%), Elphidiella brotzkajae (0.6–4.87%) and Cornuspira sp. (0.11%) (Fig. D1). A. beccarii is the most abundant taxon in this core and Elphidium shochniae and Cornuspira sp. are the rarest.

The comparison of foraminiferal abundance between the four fractions shows a maximum abundance in the 63–125 and 125–250 μm fractions. The maximal abundance of foraminifera is at 24 cm depth, with assemblages displaying a dominance by A. beccarii. The lowest foraminiferal concentration is recorded at 7 cm depth (Fig. D1). In the entire core, A. beccarii maximum abundance is at 14 cm depth and minimum abundance at the 34 cm of core, but the abundance of E. littorale is reversed at these depths.

The core US01 is characterised by silty sediments with sand and clay with low organic matter except below 27 cm depth where it reaches 5%.

Boomer et al., 2003, S. Leroy, snapshot 1999b

KBG08-01b

n/a: not available.

Kazanci et al., 2004. Leroy et al., 2006.
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