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Snapshot observation of physical structure and stratification in deep-water of the South Caspian Sea (western part)

P. Ghaffari[1], **H. A. Lahijani**[2], **and J. Azizpour**[1]

[1]Department of Physical Oceanography, Iranian National Center for Oceanography, No. 3, Western Fatemi, 1411813389, Tehran, Iran
[2]Department of Geology, Iranian National Center for Oceanography, No. 3, Western Fatemi, 1411813389, Tehran, Iran

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Correspondence to: P. Ghaffari (ghaffari@inco.ac.ir)

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Abstract

In this article, we describe physical parameters structures and different water masses using CTD measurements in southwestern part of the Caspian Sea (CS) in adjacent to Anzali Port (AP). CTD profilings were conducted along a transect perpendicular to the coastline over 13 stations from the coast down to 720 m in winter 2008. According to the results the continental shelf waters are located in surface mixed layer. Surface mixed layer extends itself down to almost 100 m in outer areas of the continental shelf with a weak seasonal thermocline layer between 80 to 140 m. Freshwaters inflow of local rivers is clearly seen outside continental shelf at the surface layers. Investigating the dissolved oxygen reveals that winter convection is traceable down to 500 m in the lateral waters over the shelf break. Among the deeper stations that are located in continental rise and abyssal plain, 350 m seems to be threshold for penetration of seasonal changes; therefore deeper waters tend to be impermeable against seasonal variances. Despite to the small variations, stability is positive in most region of the study area and temperature plays an important role in static stability and in triggering the lateral mixing. In view of both temperature-salinity and temperature-oxygen distributions in the southwestern part of the CS, three different water masses are separable in cold phase. Snapshot observation of physical properties in the early winter 2008, to some extent revealed that a mixing was triggered at least in the lateral waters of the study area.

1 Introduction

The Caspian Sea (CS) is the world's largest lake in both area and volume that lies in among Iran, Azerbaijan, Russia, Kazakhstan, and Turkmenistan (Safarov et al., 2008; Zaker et al., 2007). In addition to northern shallow basin, the CS consists of two deep basins in south and central parts which separated by a sill. Almost 60% of the total water mass of the CS is located in deep water (below 200 m) and it can be classified as a deep inland sea based on its peculiarities of the thermohaline structure and water circulation (Lebedev and Kostianoy, 2006; Peeters et al., 2000) .

The main inflow comes from the Volga River, which discharges into the northern basin, contributing about 80% of the total water inflow while freshwater input from Iranian side is extremely limited (Rodionov, 1994). Northern part of the CS plays an important role in the water circulation among the whole basin. Every cold phase the surface ice sheets covers this part and forms dense water which sinks down toward the central basin, consequently the central basin dense water flows to the southern part. Obviously the dense water formation procedure in the northern basin is depending to the freshwater discharge. Higher freshwater budget means low dense water formation and consequently causes to hampering of mixing process. Another well known scenario for mixing in the CS is winter convection. Usually winter cooling triggers mixing over the water column in the most parts of the CS. Moreover cold and dense water tongues are formed by cold shallow water of the lagoons and rivers in the lateral boundary of the basin (Terziev et al., 1992). Unfortunately this water body still is very poor in terms of instrumental observation particularly in the southern part along the Iranian coast. Therefore, it would be necessary to carry out research cruises particularly in span of the winter season when deep water formation is expected to occur (Froehlich et al., 1999).

This paper presents the results of the CTD-DO probe measurements in the southwestern part of the CS across one transect adjacent to the Anzali Port (AP), in early winter 2008. Probe profilings were exceeded down to 720 m in the Iranian waters which brought some insights to the deep water structure and winter convection in the study area. Also vertical structures of the physical characteristics and stratifications were investigated in mid and deep water of the southwestern part of the CS. Furthermore the gathered data sets revealed some new aspects of different water masses in the study area. The field survey was carried out coincide with unprecedentedly cold winter during past 40 years which reported by the local meteorological stations.

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2 Field data and CTD adaptation

2.1 Study area

The study area is located in the southwestern part of the CS adjacent to the AP (Fig. 1). Depth from the coast increases gently to about 50 m near the shelf break at almost 10 km from the coastline, after that the depth sharply increases to 800 m in 25 km. According to the bathymetric features, the continental shelf spreads from coast up to about 10 km seaward with mild declination, which ensues by a drastic depression (the continental break) extends to 20 km seaward and hits its lowest point at 600 m. The continental rise commences from this point and goes down with a relatively mild slope respect to the continental break toward the abyssal plain (top left Fig. 5). The Anzali Lagoon (AL) is located at southwestern corner of the CS beside the delta of the Sepidrood River that supplies near to 50% of riverine freshwater inflow and approximately 90% of sediment discharge to the CS along the Iranian coast, where Iranian total freshwater inflow does not exceed more than 5% of total riverine freshwater input to the CS (Lahijani et al., 2008). Moreover the area benefits from considerable precipitation and does not have a dry season (Sharifi, 2006). Therefore, to some extent the freshwater budgets of the AL and the Sepidrood River are considerable in the region. Depending on the freshwater budget, the southern part of the CS has the maximum basin wide salinity (Millero et al., 2008) and its magnitude increases toward south and east. The southern coast of the CS has a subtropical climate characterized by warm summers and mild winters (Kosarev and Yablonskaya, 1994). The air temperature is maximum in August and minimum in January (Kaplin, 1995; Kosarev and Yablonskaya, 1994). While the surface temperature attains its annual minimum of about 7°C in February, in span of warmer phase the maximum surface temperature exceeds 27°C in the southern basin (Kaplin, 1995; Tuzhilkin and Kosarev, 2005). Temperature in the deep water of the CS is well below the annual mean temperature of the seasonal mixed layer and in the southern basin it is even below the annual minimum of the monthly surface temperature (Tuzhilkin and Kosarev, 2005).

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2.2 Field measurements

A total of 13 profiles along a transect extending from the mouth of the AL towards the offshore region, as shown in Fig. 1, were carried out in the early winter of 2008. The stations were selected based on the bathymetric contours in approximately 3 km intervals to cover some specific depths along the study area. The first station located at latitude N 37°29.185′ and longitude E 49°27.959′ off AP and the last one located at latitude N 37°44.121′ and longitude E 49°40.470′ which hits its lowest point at 720 m depth. According to the bathymetric features, the profiled stations well covered the deepest parts of the Iranian waters and could be representative for deep water characteristics in this region.

Measurements of in situ physical parameters such as pressure, temperature, conductivity and dissolved oxygen were made using an Ocean Seven 316 CTD probe produced by Idronaut Italy. The probe passed pre-calibrations and lab procedures former to survey in order to achieve precise estimates of conductivity (±0.003 ms cm^{-1}), temperature (±0.003°C), and pressure (0.05% full scale) in deep water. Dissolved oxygen measurements were made using a routinely calibrated polarographic oxygen sensor which is insensitive to nitrogen, nitrous oxide, carbon dioxide and other gases. In addition to aforementioned parameters, the probe was equipped with ancillary sensors such as Chlorophyll A, pH and turbidity which is not in scope of present paper. The CTD speed into the water was maintained nearly constant at 1 ms^{-1} during casts (Pinot et al., 1997). The data quality was ensured through careful data inspection, procedural blank measurements, de-spiking and checking for duplicate records.

2.3 CTD data adaptation

Linear relationship between hydrostatic pressure and geometric depth enable us to derive depth of profiling instruments from measured hydrostatic pressure. The relationship is such that the “pressure expressed in decibars is nearly the same as the numerical value of the depth expressed in meters” (Sverdrup et al., 1942). The cause

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for the slight difference between pressure in decibars and depth in meters is found in the well-known hydrostatic relation. In consequent, different values of density give a slightly different pressure versus depth relation (Emery and Thomson, 2004). Using the average density of the CS ($\rho \approx 1.011 \times 10^3\,\mathrm{kg\,m^{-3}}$), which is slightly less than world ocean density (Tuzhilkin and Kosarev, 2004), yields a slight different pressure to depth correction coefficient in this water body;

$$\text{Pressure} \approx 0.991719\ \text{Depth} \tag{1}$$

On the other hand since the CS, is leveled approximately −27 m (with respect to the Baltic altitude system) below the measured World Sea level (Lebedev and Kostianoy, 2008), atmospheric pressure correction is inevitable in order to maintain accurate quantities of depth from recorded pressure data.

In oceanographic studies the calculation of salinity and density from CTD measurements is usually based on the empirical equations given by UNESCO (1981a, b). In view of the fact that empirical relations are valid only for water which has the ionic composition of standard sea water, they cannot be applied directly to the CS water. Millero and Chetirkin (1980) have conducted laboratory experiments to provide an empirical equation of state for the CS water. Unfortunately, the relation was valid only at surface pressure in specific temperature range. Peeters et al. (2000), using basin-wide water samples collected in the CS (IAEA, 1996), modified the experimental equations for salinity and density based on the ionic composition of the CS. Their results showed that the salinity computed from CTD data by using the standard processes of UNESCO (1981b) is systematically less than the salinity obtained from the chemical methods, and the best agreement between the instrumental UNESCO formula and the chemical titration method in the CS can be obtained by applying a modification factor of 1.1017 as follows:

$$\mathrm{Sal}_{\mathrm{casp}} = 1.1017\,\mathrm{Sal}_{\mathrm{sea}} \tag{2}$$

where 1.1017 is the correction factor, $\mathrm{Sal}_{\mathrm{sea}}$ is computed salinity using UNESCO equations, and $\mathrm{Sal}_{\mathrm{casp}}$ is the corrected salinity for the CS water. The Eq. (2) is independent of pressure, temperature and salinity ranges and could be used for different conditions and depths.

The international equation of state, i.e. density as function of temperature, salinity and pressure, for standard sea water ρ_{sea} (UNESCO, 1981a) can be applicable to the CS by using Peeters et al. (2000) approach as follows:

$$\rho\text{solution} = \rho_{\text{sea}}(T,0,P) + f(T,P)(\rho_{\text{sea}}(T,S,P) - \rho_{\text{sea}}(T,0,P)) \quad (3)$$

where in Eq. (3), $\rho_{\text{sea}}(T,0,P)$ is the density of the water computed from UNESCO formula using in situ temperature and pressure, in freshwater, $\rho_{\text{sea}}(T,0,P)$ is the density of water from UNESCO formula using in situ temperature, pressure and salinity and $f(T,P)$ is the correction factor ($f = 1.0834$), that was assumed to be constant in all computations. In the present study we computed salinity and density of the CS based on the equations presented by Peeters et al. (2000) which seemingly are applicable for whole basin concerning the basin wide supporting data in derivation.

3 Result and discussion

3.1 Physical properties structures

Figure 2a shows the temperature structure along the transect perpendicular to the coastline up to about 32 km seaward. The depth in this transect started from 5 m at the first station and extended to 720 m at the last one. Relatively shallow waters of the continental shelf which are extended to about 10 km from the coast clearly are under the influence of the AL inflow and winter cooling, therefore winter convection penetrate almost down to seabed and vertical mixing is evident in this region. Contrary to existing the summer strong thermocline in southern part of the CS that partially maintain itself in continental shelf zone (Zaker et al., 2007), water body in the cold phase over the continental shelf is completely ventilated and located in surface mixed layer. However the inflow of the AL is traceable over whole continental shelf zone. The cold water

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inflow of the AL (approx. 1°C and not shown in Fig. 2a due to resolution), warms up to 9°C by dilation procedures with ambient waters as expands seaward. Owing to the cold phase erosion, the seasonal thermocline inevitably sinks down to the deeper layers and is located between 80 to 140 m depth with 2°C difference. Moving shoreward from the outer stations the deepening of thermocline layer is more evident (Fig. 2a gray band). Therefore a well mixed layer with approx. 100 m thickness is formed over the thermocline layer in cold phase. According to Bruevich (1973), the maximum depth of winter convection in deep part of south CS and near Iranian coastal zone differed between 80–100 m and 40–60 m respectively, which apparently is exceeded during this specific winter. Below the thermocline layer the temperature decreases gradually to 6.5°C at about 350 m in the lateral boundary, and follows the same trend toward the seabed. Therefore based on the isotherm patterns, the seasonal mixing does not penetrate further than 350 m in the study area, and deep waters (lower than 350 m) behave almost impermeable against seasonal variances.

Vertical structure of salinity in the study area in early winter 2008 is presented in Fig. 2b. Outside of the continental shelf zone, the variation of the salinity is very small and its difference from surface to 300 m is about 0.2‰, i.e. at surface layer salinity is 12.242‰ and reaches 12.425‰ at almost 350 m and maintains approximately the same value to the seabed. The value of the measured salinity in the study area is within well agreement with previously reported observations (Kaplin, 1995; Kosarev and Yablonskaya, 1994; Tuzhilkin and Kosarev, 2005). The radical transformation of the vertical salinity structure of the CS waters is the most prominent event in the hydrological condition of this water body. By plunging the water salinity over the whole Caspian basin, in the early 1970s, a stable vertical salinity stratification was formed (Tuzhilkin and Kosarev, 2005). This stratification was most pronounced in the early 1990s and in that period, vertical differences in salinity values made 0.5‰ in the southern part of the CS. However lately, a reverse trend has begun to manifest itself and leads again to quasi-homogeneous vertical distribution of salinity values (Tuzhilkin and Kosarev, 2004). Although a slight stratification is distinguishable in the salinity structure

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across the water column in the outer shelf zone (Fig. 2b), but it can be mentioned that southwestern part of the CS is quasi-homogenous in terms of the salinity.

Horizontal variations of salinity over the continental shelf represent the influence and extend of freshwater inflow from the AL. Horizontal differences in salinity values among the continental shelf are noticeable and vary in range of 1.9‰ (not shown in Fig. 2b due to resolution) to 11.708‰. The plume of AL inflow under the effect of negative buoyancy goes down to the level where the density of the plume and surrounding waters are equal. Moreover the convection can occur due to the heavy suspended loads in the AL inflow. The external freshwater budget, mainly river runoff, is the evident cause of the evolution in the salinity structure of the study area. Albeit the salinity growth with depth is rather small, but concerning to the present vertical distribution in southwestern part of the CS, the structure could still be classified as the subarctic type, which has been observed from middle 1990s (Tuzhilkin et al., 2005).

Density profiles are a function of both salinity and temperature. Apparent density is calculated using the adopted equation of the state for the CS water (Fig. 2c). The vertical structure and evolution of the density field represent a function of its physical parameters (temperature and salinity); the function depends on the seasonal variability of these parameters. Accordingly, the density in the upper layer slowly increases (from 1009.1 to 1009.6 kg m^{-3}) down to almost 80 m depth in the cooling phase. A quasi-homogeneous layer about 80 m thick is formed in the density structure and a broad seasonal pycnocline is formed under this layer which is extended to 140 m depth resemblance to thermocline layer. Contrary to the density field in the warming-up phase (Zaker et al., 2007), in span of heat loss phase in the southern CS, the temperature drop results an increase in the upper layer density and the thickness of the mixed quasi-homogeneous layer.

Variances of the pycnocline occur in accordance with the seasonal evolution of the density field. The density field in the deep waters associates with temperature shrinkage and rather salinity growth. Underneath the thermocline layer, density is 1010.5 kg m^{-3} and with a small gradient increases to about 1013.0 kg m^{-3} at seabed.

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Although the vertical structure of the density closely mimic the salinity and temperature structures, but mostly temperature has a greater impact on the density field rather than salinity, in most world oceanic bodies (Gill, 1982; Stewart, 2006). Considering the slight variations of salinity in the study area respect to the variations of temperature, the density structure is highly correlated with the temperature field. Horizontal variations of density at the mouth of the AL and partly in the continental shelf area could be attributed to the intrusion of the AL inflow which adds riverine water into the sea.

The AL low dissolved oxygen inflow water (approx. 5.0 mg l^{-1}) is ventilated and mixed over the continental shelf area. The dissolved oxygen concentrations reach its highest value 12.0 mg l^{-1} on the middle of continental shelf in early winter 2008 (Fig. 2d). The well ventilated and highly oxygenated water is extended through the surface mixed layer over the continental shelf and shelf break (approx. 20 km from coastline). The oxygen concentrations in the upper 100 m mixed layer, owing to high ventilation and low temperature, is high and mainly ranged between 10.0 to 7.0 mg l^{-1}. There is a considerable depression about 100 m in the dissolved oxygen isoclines over the continental break and continental rise. The minimum oxygen layer from 400 m in outer area is shriveled below the 500 m level in the vicinity of the shelf break along the lateral boundary. The declination of the oxygen contours over the shelf break could be qualified by winter convection and sinking upper waters which pushing down the oxygen isoclines (Fig. 2d). The oxygen concentrations are about 6.5–7.5 mg l^{-1} at 100 m surface layer and the mean annual absolute content of dissolved oxygen decreases with depth from 4.0 to 2.5 mg l^{-1} in the 200–600 m layer of the south CS (Tuzhilkin et al., 2005). Locating the surface layers oxygen (6.0 mg l^{-1}) in 200 m is indicating that fully ventilated continental shelf waters spread up over the continental shelf break and lean the oxygen isoclines under the influence of the winter convection. The tilt of oxygen isoclines in lower layers shows the penetration of high oxygenated waters from the upper layers; i.e. above 200 m. Because biodegradation leads to oxygen depletion, the intrusions are most likely recent features. Although a major convection event has not occurred such as what happened in 1976, which caused oxygen concentrations to

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reach around 9.0 mg l^{-1} in the deep waters of the central basin (Kosarev and Yablonskaya, 1994), but it could be mentioned that in the early winter 2008, a considerable winter mixing triggered at least in the lateral water boundary of the southwestern part of the CS; which could be traceable down to 500 m depth in terms of dissolved oxygen. At the deepest points of the study area, dissolved oxygen concentrations are strongly under-saturated due to oxygen consumption and reach less than 0.5 mg l^{-1} (in account to resolution not shown in the Fig. 2d), which almost is one third (1.5 mg l^{-1}) of the near-bottom oxygen concentration averaged over 1958–2000 (Tuzhilkin et al., 2005). This decrease is caused by the weakening of the winter time convective ventilation in the deep waters and strengthening of the abyssal destruction processes.

3.2 Stability and water masses

In terms of the measured temperature and salinity profiles, one can calculate the Brunt Väisälä frequency N^2, which is representative for the local stability (the stability with respect to an infinitesimal vertical displacement of water parcel) across the water column (Gill, 1982; Sverdrup et al., 1942):

$$N^2 = N_T^2 + N_S^2 = g\alpha(\frac{dT}{dz} + \Gamma) - g\beta\frac{dS}{dz}, \quad (4)$$

where N^2 is the sum of the temperature-dependent stability, N_T^2, the salinity-dependent stability, N_S^2, (Fig. 3), g is the gravitational acceleration, α is the thermal expansivity, β is the haline contraction coefficient, Γ is the adiabatic lapse rate, and z is the depth (positive upward). Our total stability and temperature-salinity dependent stabilities calculations were based on the data of the outer stations which located beyond the continental shelf. All data were averaged in 10 m depth bins (Vollmer et al., 2002) and smoothed out utilizing the linear least-squares method.

According to the N^2 profiles in the last station which is the deepest one in the study area, small scale instability is evident at surface layer. The process of the mixing is

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account for small vertical scale static instability in near the surface and interfaces between different waters (Emery et al., 2006). N_T^2 and N_S^2 values are positive in the most regions of the study area which confirms the previous results (Peeters et al., 2000). In the deep water region of the southwestern part of the CS, the contribution to the stability by the temperature gradient is significantly larger than that by the salinity gradient, because salinity variations are small and water column is quasi-homogeneous in terms of salinity. The maximum quantity for N^2 appears in the thermocline layer which getting deeper in the shallow stations. Therefore, a considerable slope in thermocline layer toward the sea (from the station 7 to the station 13) is evident in the study area. Below 350 m the stability decreases gently and becomes zero in last 20 m. Heating and cooling of the water column during warm and cold phases are account for partial or complete turnover in lakes which often supported by wind forcing (Hohmann et al., 1997). Deepening of the thermocline layer from greater depths towards the shelf break and shallower regions could support the hypothesis that turbulent diffusion cannot be the only mixing process. Hence, the processes causing oxygen-rich water to advect as density plumes from shallower parts down to the greater depths of the study area must be driven by horizontal temperature differences at shallower depths. Most probably the continental shelf water well cooled by AL inflow, extreme winter and horizontal mixing of such water with moderately warmer shelf break water creates intense thermal bars which can produce water dense enough to sink. Salinity differences most likely do not play a significant role in driving the mixing process, in particular because salinity within the inflow is lower than that of the ambient water. Although the inflow of the AL is considerably fresher than the CS water, which could prevent such a mixture from sinking, but convection could also be generated by an increased load of suspended particles in the water as is the case for riverine inflows (Peeters et al., 2000; Vollmer et al., 2002). In the case of the CS, salinity of the inflows is lower than the ambient waters, so most probably the suspended loads play a substantial role in the mixing processes.

In order to index of water structures (temperature and salinity) and separation various water types and masses in the study area, T-S and T-O diagrams are presented in

Figs. 4 and 5. As it is demonstrated in Fig. 4a, three different water masses in deep part of study area (stations 07 to 13) are separable. The surface layer extends from surface to 50 m whereas temperature and salinity are almost constant along this layer. Another water mass is distinguishable between 50 to 150 m with rather significant decrease in terms of temperature (about 3°C), and to some extent a constant trend in terms of salinity. The third water mass located from 150 m to seabed which 0.2‰ increase in salinity and 1°C decrease in temperature. Deep layer (150 m to seabed) is the portion of the temperature versus salinity diagram which shows less scatter, and forms the water mass for the main sub-thermocline which is characterized by quasi-horizontal isotherms as observed in Figs. 2 (a and b).

In Fig. 4b, T-S diagram for the stations 01–13 is plotted against station 03, of IAEA cruise (in the southern basin of the CS) in winter 1996. The coastal and continental shelf areas (stations 01–06), construct completely different water with low salinity and temperature quantities under the influence of the freshwater inflows, however surface temperature in the deeper stations is about 10°C. The surface temperature in the southern basin attains its annual minimum of about 7°C, when the basin-mean surface temperature is about 9°C (Kosarev and Tuzhylkin, 1995). Although in very cold wintertime water vertical rotation is effective in deep water formation (Kosarev, 1962; Terziev et al., 1992; Terziev, 1992) and according to Fig. 4b the lateral water temperature are well below the annual minimum and capable to sink to deeper layers, but the increase in density due to seasonal cooling in deeper stations is not sufficient to cause convection to the deepest region within the study area.

In view of the fact that distribution of salinity is semi-homogeneous in vertical outline in the CS, T-O diagram is more illustrative to identify water masses in this water body (Terziev et al., 1992). According the Fig. 5 dissolved oxygen ranges between 4.5–6.0 mg l^{-1} in stations 01 and 02 which indicating cold and low dissolved oxygen inflow waters of the AL. Additionally over the continental shelf area owing to ventilation processes, dissolved oxygen concentrations increase and reach to the maximum amount near to 12.0 mg l^{-1} at the surface layer. At the mean time temperature increases gradually from 5 to 10°C as proceeding seaward over the continental shelf, therefore there are two distinguishable waters, which completely detached from the deeper stations. In terms of oxygen concentrations, the upper layer in the continental shelf area is under the influence of fresh and cold water inflow of the AL, and lower layer adjoined with the main southwestern Caspian waters.

Among the deeper stations (from station 05 to 13), T-O diagram illustrates three water masses: surface, middle and deep waters. The surface layer extends from surface to 50 m. In these stations the maximum dissolve oxygen concentrations were observed in surface layer about 9.0–10.4 and 6.0–7.0 mg l^{-1} over the shelf break (stations 05–10) and continental shelf rise (stations 11–13) respectively, where the temperature shows its maximum value (averagely 10°C). Beneath the surface layer, middle layer extends from 50–150 m which still the shelf break waters are warmer and more oxygenated than the seaward waters. Below the middle layer, the dissolved oxygen decreases drastically from almost 6.0 mg l^{-1} to 0.5 mg l^{-1} which forms the deep waters. Although Terziev et al. (1992) results stated that only two water masses, surface (surface to 50 m) and deep (from 50 m to seabed) are hardly separable in the cold phase in the south CS, but our measurements in early winter 2008, confirms three clearly distinguishable different water masses in terms of temperature-oxygen in the study area. In deep layer there is a considerable difference in oxygen level, but temperature shows the same values along the isobars. The separation of T-O trends in the shelf break and seaward waters could be attributed to lateral winter mixing in the study area which causes penetrating high oxygenated waters to lower depths over the shelf break. As it illustrated in Fig. 2a and d, the isotherms and oxygen contours were pressed and bent down over the shelf break zone under the influence of winter convection, therefore vertical displacement of water is occurred in this area which cause partitioning of T-O trends. Additionally separation can lead us to trace the lateral winter mixing down to approximately 500 m in the shelf break zone.

4 Conclusion remarks

Physical (temperature, salinity, density) structures and dissolved oxygen concentrations and their variations over the southwestern part of the CS off AL, Iran were presented and discussed using field survey data collected in early winter 2008. Also we were discussed about various water masses in study area. The variability of the vertical thermohaline structures is caused by both local external factors (heat and moisture fluxes through the sea surface, the momentum and relative vorticity of the wind), and the local characteristics of water circulation. Based on the recorded data it is concluded that:

- The continental shelf area is located in well mixed layer and fresh and cold water inflow of AL is traceable over the shelf break. Outside the continental shelf, thickness of well mixed layer differs from 120 to 80 m as progressing seaward, therefore it seems deepening of thermocline layer is happened in larger scale than to former estimations in southern waters of the CS. According to the isotherm patterns, the seasonal mixing does not penetrate further than 350 m in the deeper stations (continental rise and abyssal plain areas), thus the waters (lower than 350 m) in these areas behave almost impermeable against seasonal variances.

- Although a slight stratification is distinguishable in the salinity structure across the water column which enable us to classify it as subarctic type, but it can be mentioned that southwestern part of the CS is quasi-homogenous in terms of the salinity.

- Based on the dissolved oxygen feature in the study area, it seems that fully ventilated continental shelf waters spread up to end of the continental shelf break and lean the surface oxygen contours down to the 200 m, under the influence of the winter convection. The tilt of oxygen isoclines in lower layers shows the penetration of high oxygenated waters from the upper layers which seem to be recent features. Although the deep waters in the seaward stations behave as impermeable against seasonal variations, but in terms of dissolved oxygen concentrations it could be mentioned that in the early winter 2008, a mixing triggered at least in the lateral waters of the study area; which could be traceable down to 500 m.

- It could be mentioned that stability is positive in most of the regions in the study area and mostly affected by temperature gradient than salinity gradient, since the deep water region of the southwestern part of the CS is almost homogeneous in terms of salinity.

- T-S and T-O diagrams are utilized in order to index of various water types and masses in the study area. The coastal water that is fairly located in AL inflow and continental shelf water behave like as completely detach masses from the rest of the study area. Within the deeper stations three different water masses i.e.; surface which extended from surface to 50 m, middle water mass distinguishable between 50 to 150 m, and deep water mass located from 150 m to 720 m is evident in both T-S and T-O diagrams. These results demonstrate a new frame of water masses particularly in aspect of temperature-oxygen structure in the study area. Despite to previous studies which were drawing two (surface and deep) water masses that were not completely separable, present study confirms three sharply distinguishable masses.

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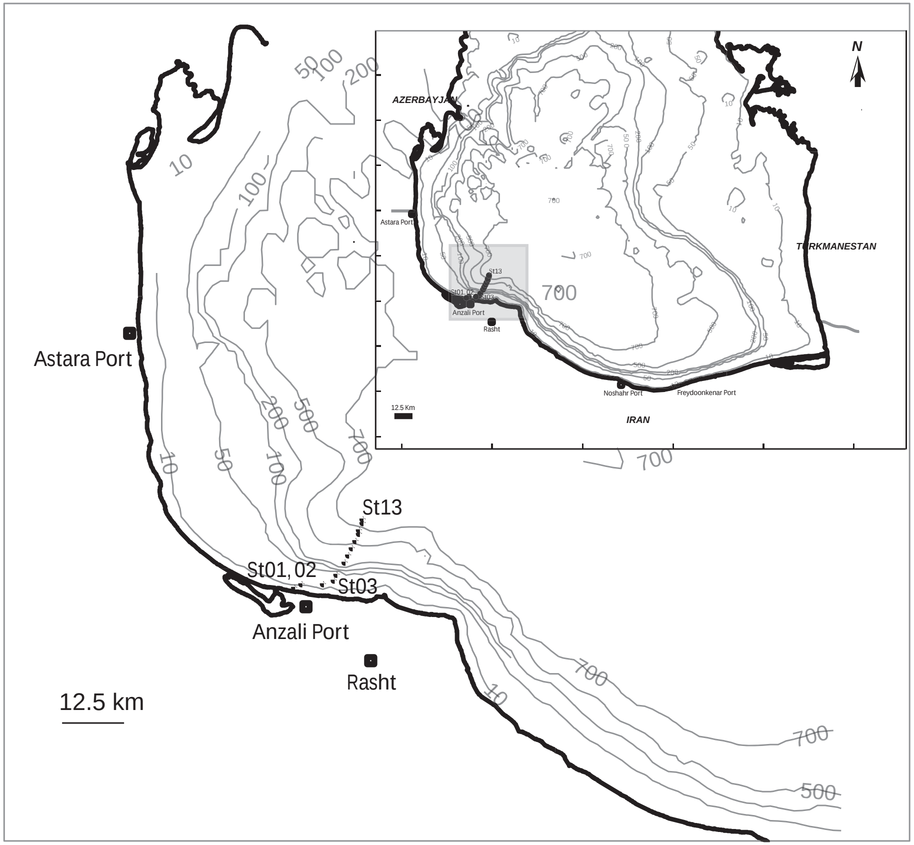

Fig. 1. The study area and CTD stations.

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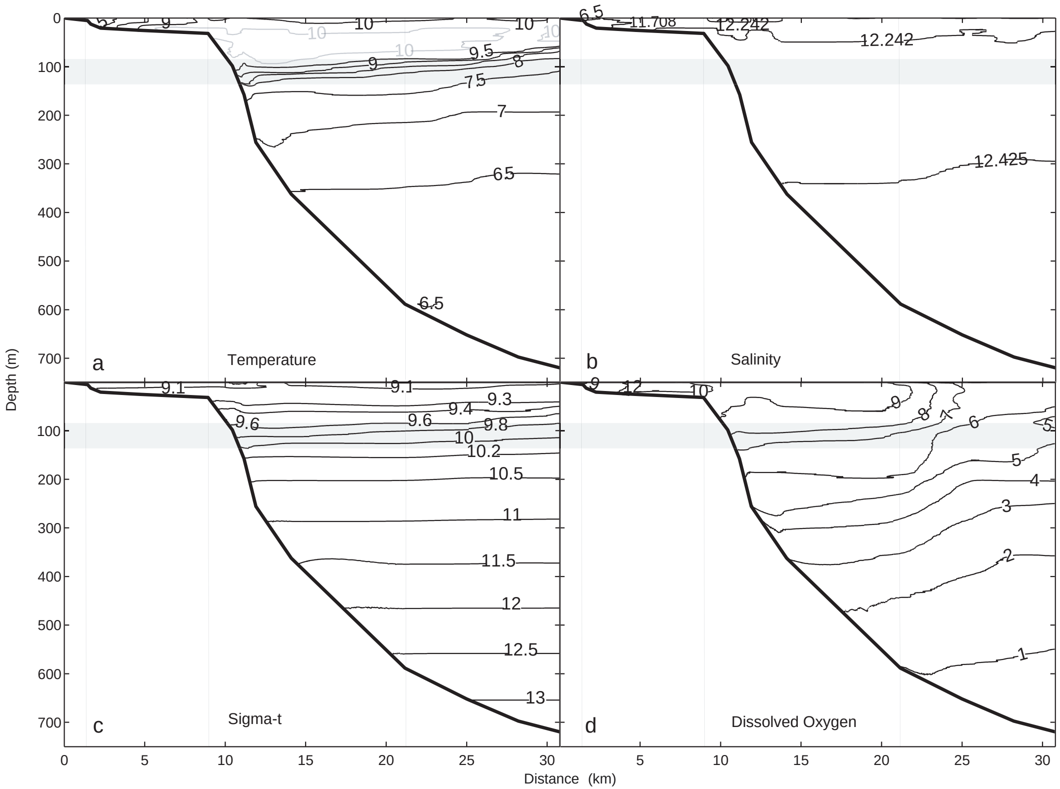

Fig. 2. Physical properties structures in cross-shore transect in the southwestern part of the CS; off AP in early winter 2008, the left side of the graphs is coast and right side is seaward (northward); vertical temperature structure **(a)**, vertical salinity structure **(b)**, vertical density distribution **(c)**, and dissolved oxygen concentrations **(d)**.

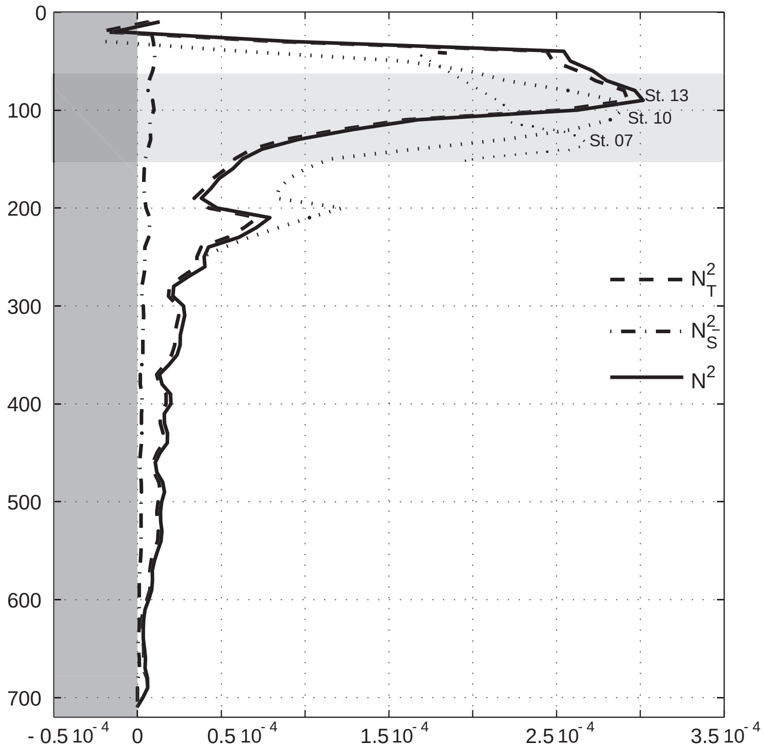

Fig. 3. Brunt Väisälä frequency N^2, which is a measure of the local stability of the water column, at the deepest station (solid line) and two shallower stations (dotted line) in the study area. N^2 is the sum of the contribution to stability by the temperature gradient N_T^2 and by salinity gradient N_S^2 (see Eq. 4).

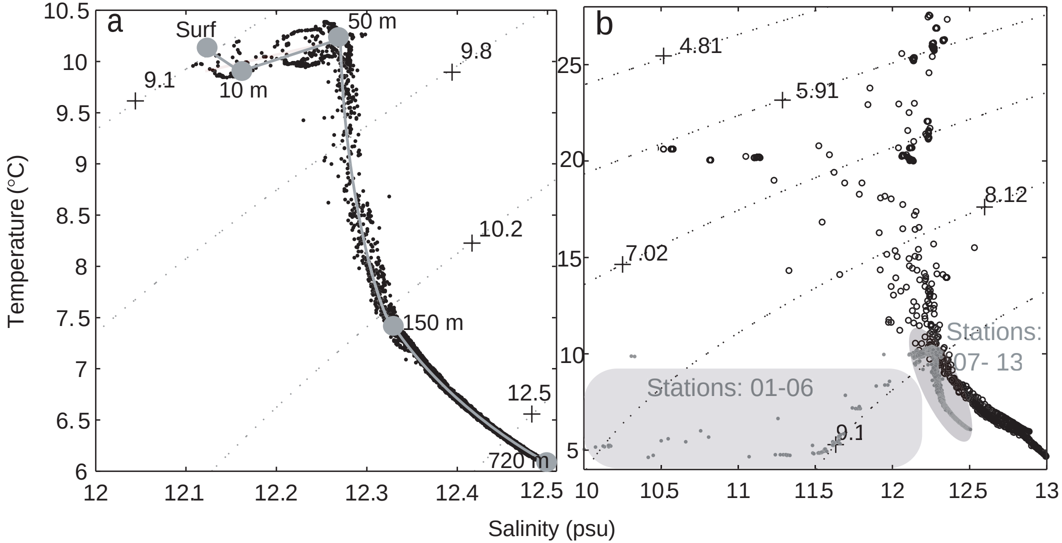

Fig. 4. T-S diagrams for outside the continental shelf (stations No. 07–13) in the southern western part of the CS in early winter 2008 **(a)**, T-S plot of station 03 of IAEA cruise 1996 which located in the south CS, in comparison with present study data **(b)**.

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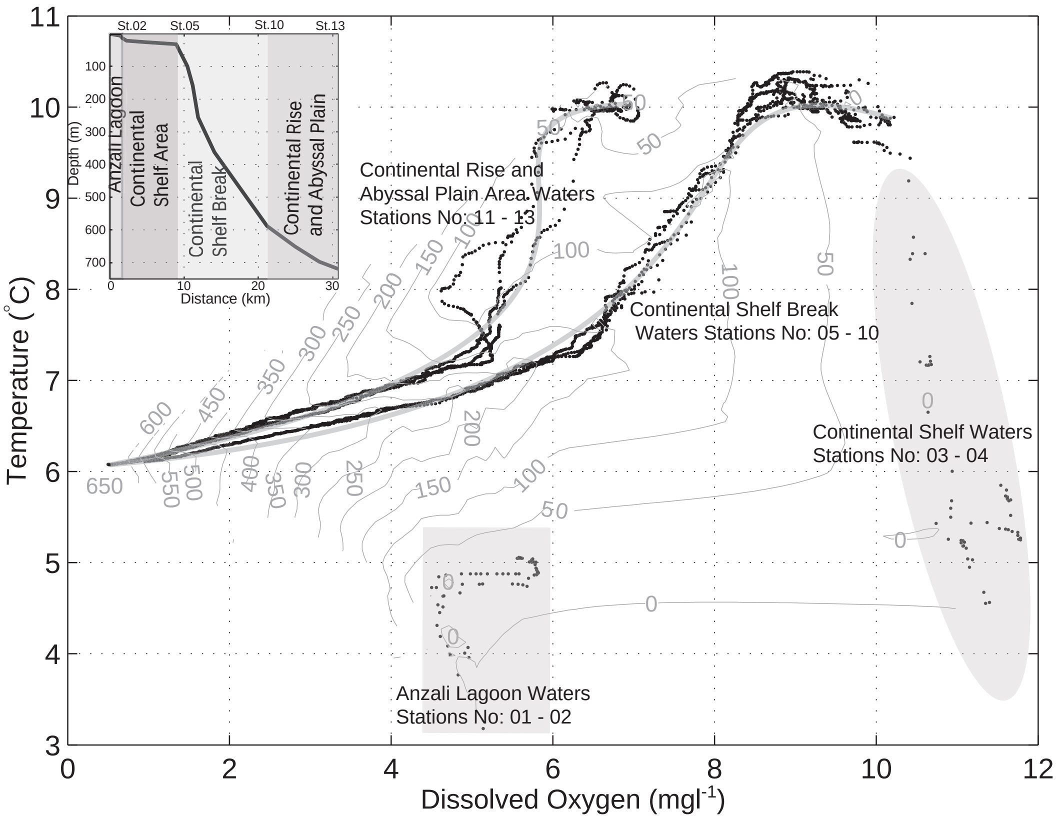

Fig. 5. T-O diagram along the transect perpendicular to the coastline in the study area in early winter 2008; the top left graph demonstrate bathymetric features of the study area, the bottom shaded areas left to right represent dissolved oxygen concentrations versus temperature in stations 01–02 AL (inflow) and stations 03–04 over the continental shelf respectively.

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