A HYCOM modeling study of the Persian Gulf: 2. Formation and export of Persian Gulf Water

Fengchao Yao[1] and William E. Johns[1]

Received 8 September 2009; revised 8 August 2010; accepted 19 August 2010; published 19 November 2010.

[1] The circulation and water mass transformation processes in the Persian Gulf and the water exchange with the Indian Ocean through the Strait of Hormuz are studied using the Hybrid Coordinate Ocean Model (HYCOM). Model results suggest that intruding Indian Ocean Surface Water (IOSW) is transformed into hypersaline waters with salinity >41 practical salinity unit by the fresh water loss in the northern end of the gulf and in the southern shallow banks. During wintertime, intense heat loss from sea to air leads to the formation of cold and saline waters both in the northern gulf and in the southern gulf. Dense waters formed in the southern gulf have higher salinity and spill into the axial deep trough of the gulf as a sporadic bottom outflow in winter, which leads to high-salinity pulses in the strait as observed, whereas in summer their buoyancy is increased due to heating and they are exported as a warm yet salty intermediate depth flow through the strait. Dense waters formed in the northern gulf propagate toward the strait along the axial trough throughout the year. Correspondingly, there are two branches of seasonal overturning circulation in density classes: the northern branch peaking in August with strength of 0.13 Sv and the southern branch peaking in February and December with strength of 0.08 Sv. These two branches outflow together feed a seasonally varying deep outflow through the strait with an annual mean volume transport of 0.12 Sv.

Citation: Yao, F., and W. E. Johns (2010), A HYCOM modeling study of the Persian Gulf: 2. Formation and export of Persian Gulf Water, *J. Geophys. Res.*, *115*, C11018, doi:10.1029/2009JC005788.

1. Introduction

[2] The Persian Gulf (referred as the gulf hereinafter) is the source region of the Persian Gulf Water (PGW), one of the most saline water masses in the world ocean. The PGW exits the gulf through the Strait of Hormuz (referred as the strait hereinafter) as a deep outflow, and descends across the shelf in the northern Gulf of Oman, entrains ambient fresher water and is intensively diluted. The PGW reaches its neutrally buoyant level at depths between 200 and 350 m and spreads laterally as a subsurface salinity maximum in the north Indian Ocean [*Rochford*, 1964; *Wyrtki*, 1973; *Bower et al.*, 2000; *Prasad et al.*, 2001]. Together with the Red Sea Water, the PGW maintains the high-salinity intermediate water inthe upper 900 m in the north Indian Ocean [*Wyrtki*, 1973].

[3] Study of the formation processes and export mechanisms the PGW is limited. *Swift and Bower* [2003] compile the available historical hydrographic data and show that the densest waters are formed in the northern gulf during winter, and that these waters propagate southward toward the strait throughout the year. Slightly less dense, though saltier, waters are formed on the shallow southern banks, which spill more intermittently into the deep gulf. However, *Kämpf and Sadrinasab*'s [2006] model study suggest that the densest water in the gulf is formed during winter in shallow banks in the southern gulf.

[4] *Reynolds*' [1993] winter and early summer surveys are the only available basin wide synoptic observations and provide valuable measurement about the PGW formation and export processes. Vertical sections of temperature, salinity, and potential density for the axial section and section F, which connects the southern shallow banks to the deep central trough, in winter and early summer are reproduced in Figures 1a, 1b, 2a, and 2b. Their locations are shown in Figure 3.

[5] The winter sections (Figures 1a and 1b) suggest that water in the gulf is cold and nearly uniform in temperature down to the bottom, resulting in a weak stratification. The two layer vertical density structure in the gulf interior is mainly controlled by the salinity. The coldest water with temperature < 16°C is found in the northern end of the gulf. As a combined result of low temperature and high-salinity, dense waters ($\sigma_\theta > 29.5$) are formed in two zones, the northern gulf and the southern banks, with the water in the latter being warmer but saltier. The dense water on the southern shallow banks, although exposed broadly to the interior, spills into the deep basin only in limited locations around section F in Figure 1b [*Swift and Bower*, 2003]. The isopycnals and isohalines in section F outcrop at the surface on the shallow banks and dip down gradually toward the deep basin without intersecting the bottom, indicating that the dense, salty water from the

[1]Division of Meteorology and Physical Oceanography, Rosenstiel School of Marine and Atmospheric Science, University of Miami, Miami, Florida, USA.

0148-0227/10/2009JC005788

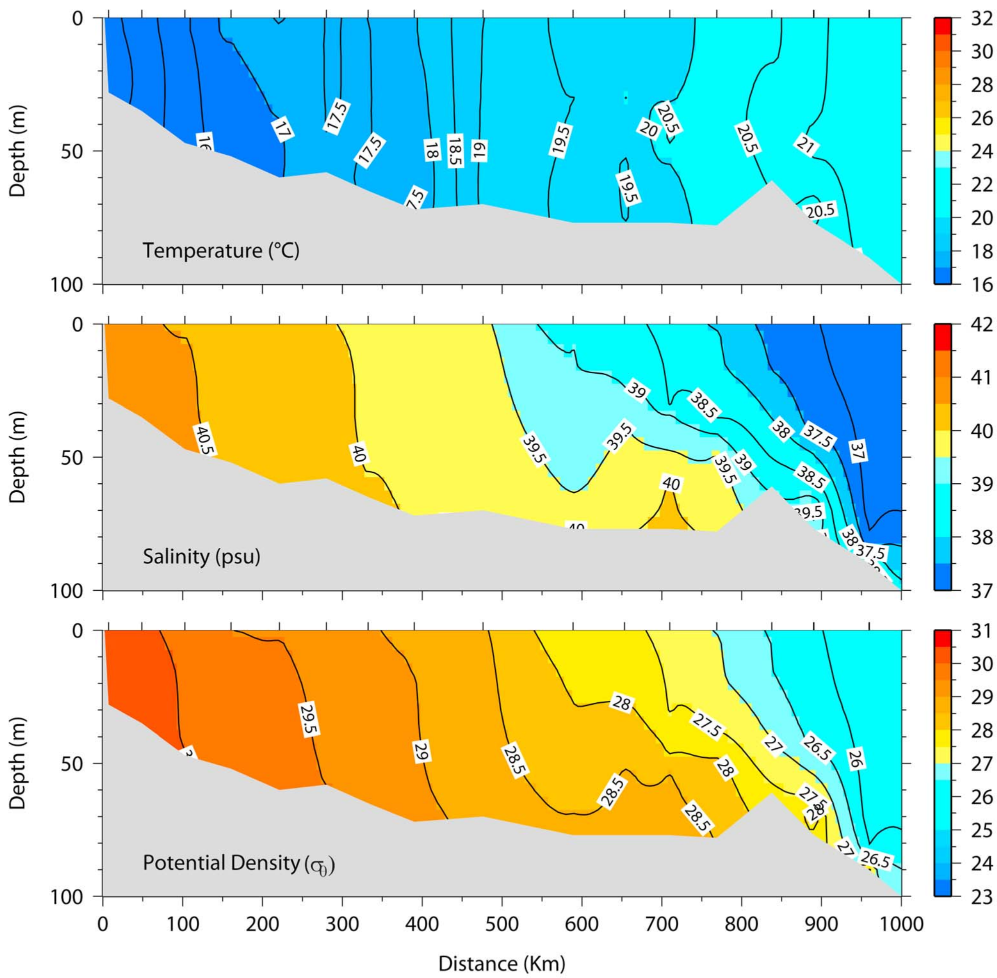

Figure 1a. Sections of temperature, salinity, and potential density along the axis during *Reynolds*' [1993] winter survey. See Figure 3 for section location.

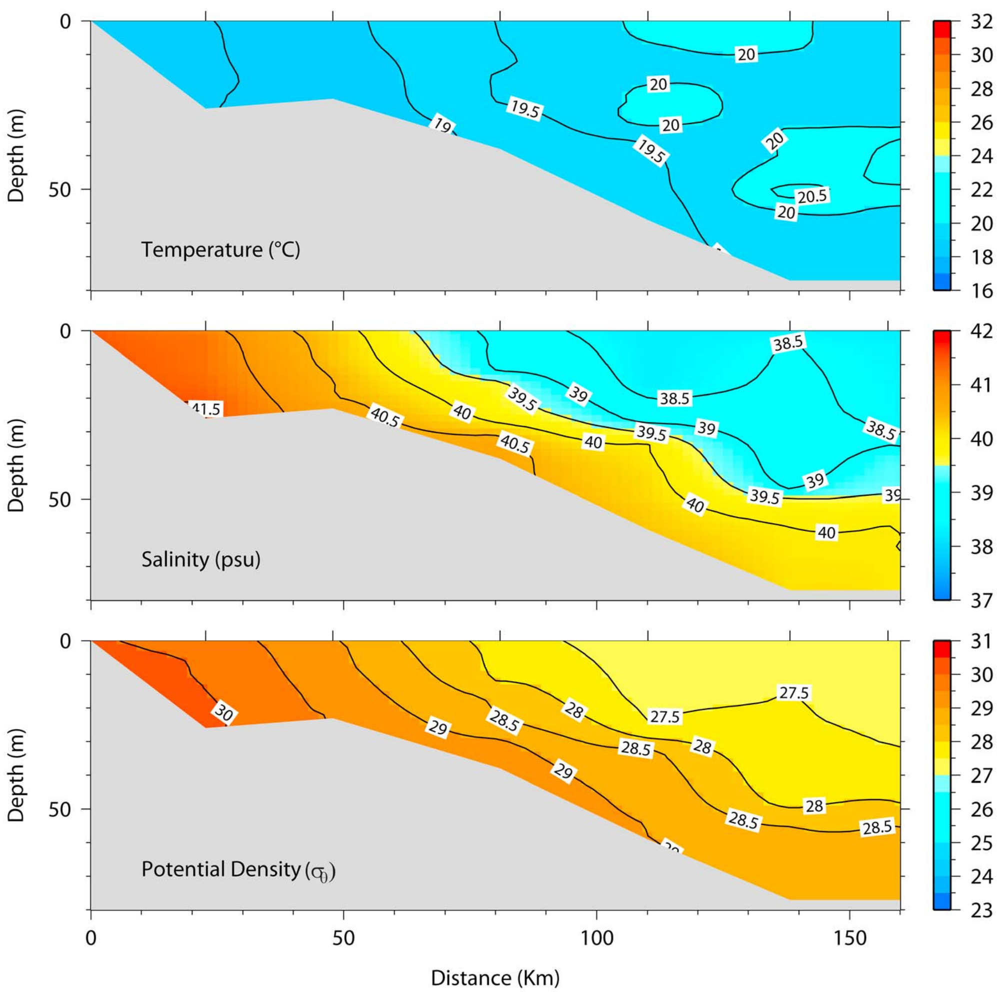

Figure 1b. Sections of temperature, salinity, and potential density along Section F during *Reynolds*' [1993] winter survey. See Figure 3 for section location.

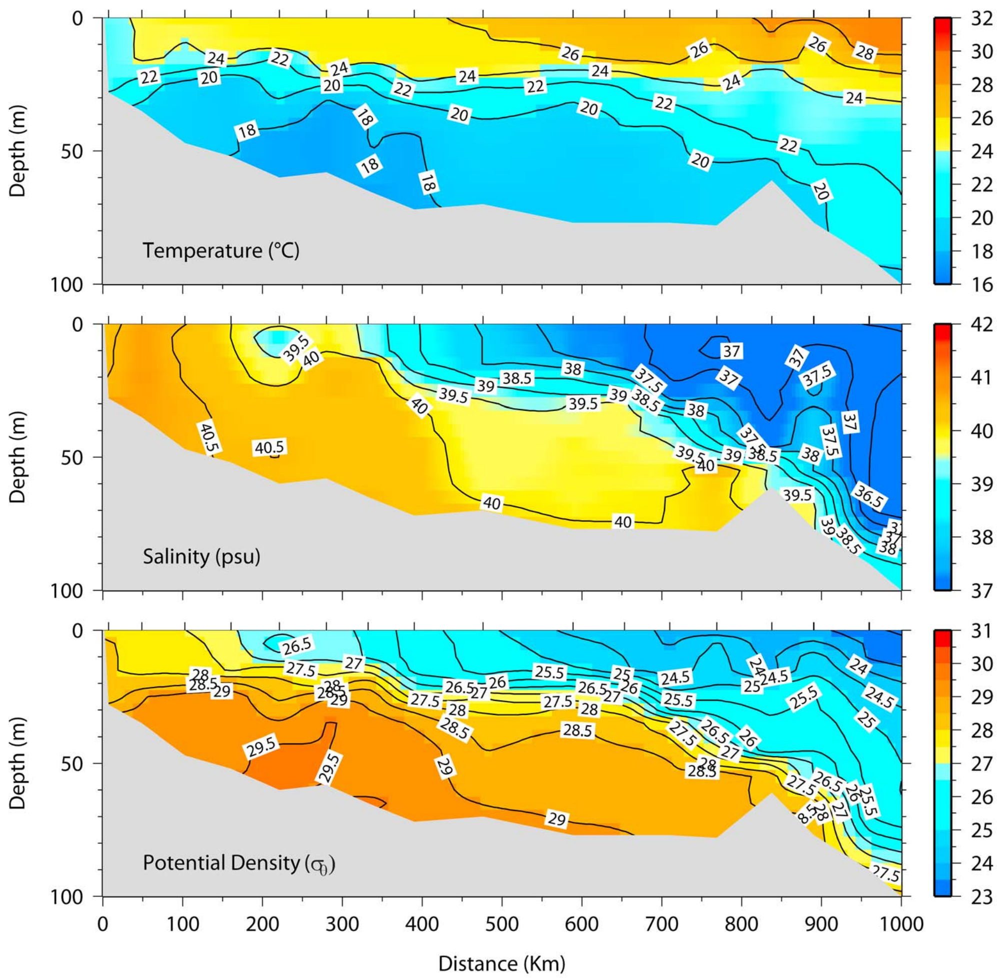

Figure 2a. Sections of temperature, salinity, and potential density along the axis during *Reynolds*' [1993] early summer survey.

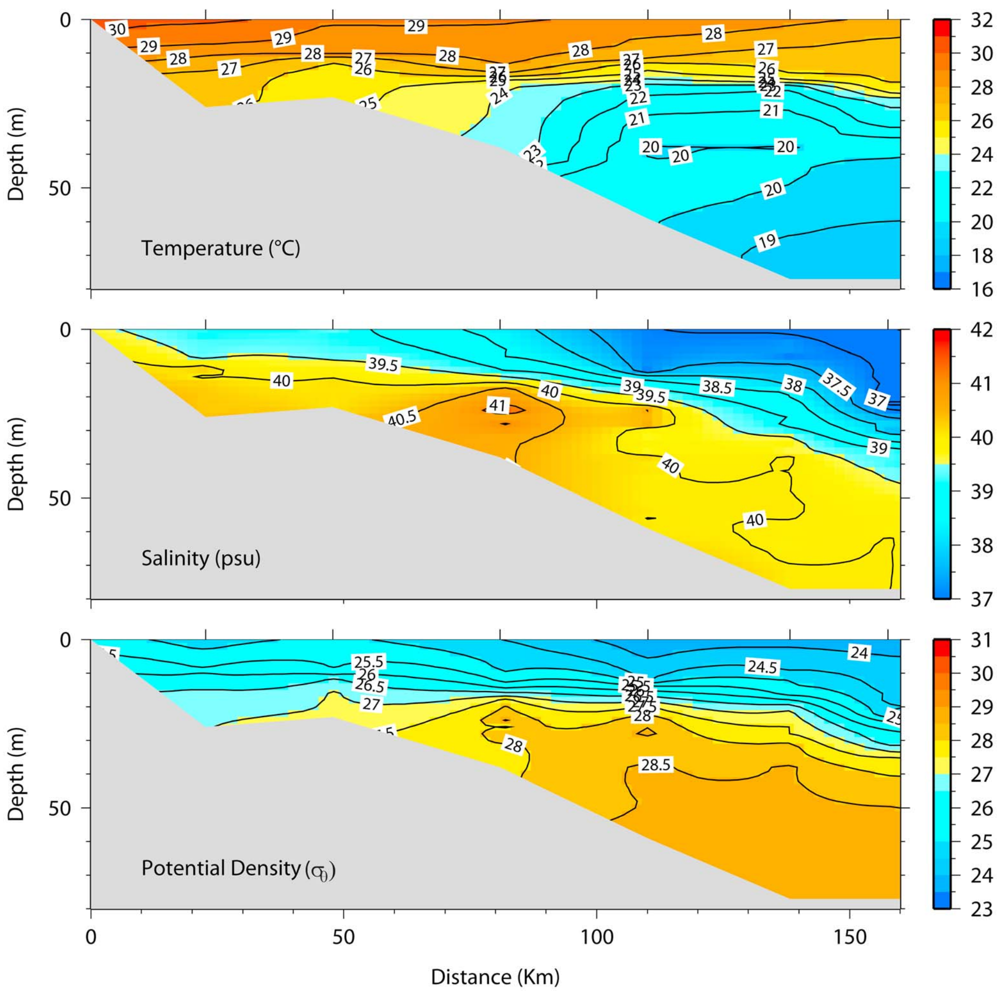

Figure 2b. Sections of temperature, salinity, and potential density along Section F during *Reynolds*' [1993] early summer survey.

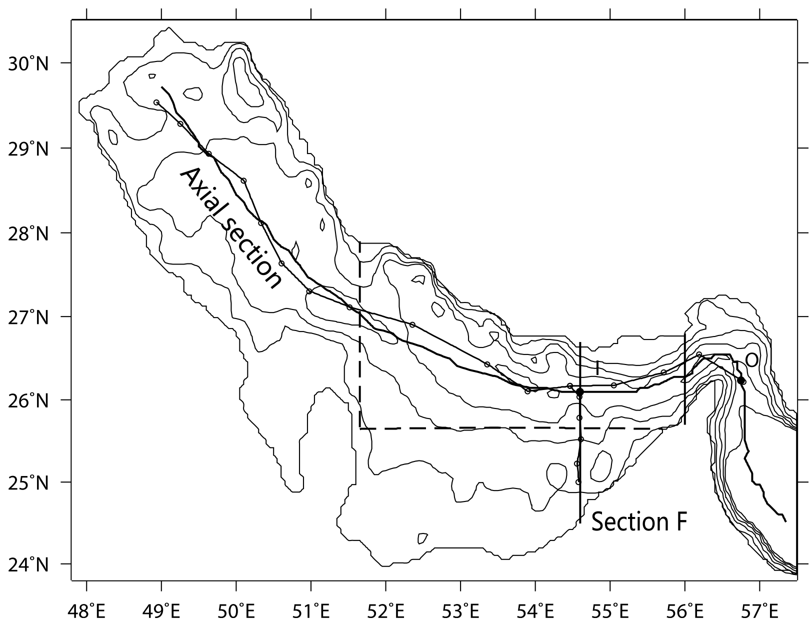

Figure 3. Locations of the sections in the model diagnosis. Three solid lines indicate the axial section, section along 54.6°E, and section across the strait along 56°E. Solid lines with circles indicate the locations of observations by *Reynolds* [1993]. The dashed lines along 51.7°E and 25.6°N set the boundaries for the northern gulf and southern banks, respectively. The two dots on the axial section denoted by "I" and "O" represent locations where the pressure field is diagnosed in Figure 15. The interval of the contours of bathometry is 20 m.

shallow banks sinks to the deep basin. Further evidence of such spilling of dense water is the bottom salinity maximum (>40 practical salinity unit (psu)) at a distance of 700 km in the axial salinity section (Figure 1a), which was also present in *Brewer and Dyrssen*'s [1985] axial section.

[6] In early summer as the surface layer is warmed up by increased insolation, a seasonal thermocline is established in depths of 20–30 m, and cold bottom water along the axial trough is isolated from the warm, fresher IOSW, which now extends much farther into the gulf [*Yao and Johns*, 2010, Figures 2a and 3]. With flatter isopycnals, the advection of the dense bottom water toward the strait from the north is more evident. The surface water on the southern banks, warmed up to more than 30°C, is not dense enough to reach the bottom in the deep gulf, and it spreads laterally along the UAE coast.

[7] The strait plays a critical role in determining the circulation in the gulf by constricting the water exchange between the gulf and the open ocean. The observations in the strait [*Johns et al.*, 2003] suggest a three-component water exchange structure consisting of (1) a relatively fresh surface inflow of IOSW on the northern side of the strait, (2) a saline deep outflow of PGW in the southern part of the strait, and (3) a mean surface outflow of intermediate salinity on the southern side of the strait which varies seasonally, including a relatively fresh inflow in spring. *Johns et al.* [2003] find the deep outflow to be relatively steady throughout the year, with annual mean volume transport of 0.15 Sv. The salinity of the outflow has a mean value of 39.3 psu and varies considerably on intraseasonal time scales, with largest fluctuations ranging from 39.5 to 40.8 psu in boreal winter.

[8] *Yao and Johns* [2010] presents simulations using Hybrid Coordinate Ocean Model (HYCOM) are shown to be successful in reproducing the observed features of the surface circulation in the gulf, and this paper focuses on the water mass formation processes forced by surface fluxes, and how the waters formed in the two different formation zones seasonally contribute the outflow in the strait.

[9] The configuration of the model, including initialization, boundary conditions, and setup of atmospheric forcing, are described in detail by *Yao and Johns* [2010]. Although the simulation under high-frequency forcing described therein produces more realistic surface temperature fields than the climatological run, both experiments reveal similar seasonal patterns and inherent physical mechanisms of the circulation, and in the following discussion the results from the climatological run are mostly used. This paper is organized as follows. In section 2, the water mass formation processes in relation to the atmospheric forcing are discussed and the pathways of the export of the dense waters in the model are presented. To further quantify the overturning circulation, the circulation in discrete density classes for each of the two formation zones is given in section 3. In section 4, the deep outflows through the strait are discussed and compared with observational results. An overall summary from the two-part study is provided in section 5.

2. Water Mass Formation

[10] The gulf is subject to constant fresh water loss throughout the year and seasonal surface heat flux (Figure 4). The net buoyancy flux over the gulf is dominated by the surface heat flux and experiences a strong seasonal cycle with net buoyancy loss in winter and net buoyancy gain in summer. The processes of the dense water formation, especially in response to the seasonal surface heat fluxes, and its propa-

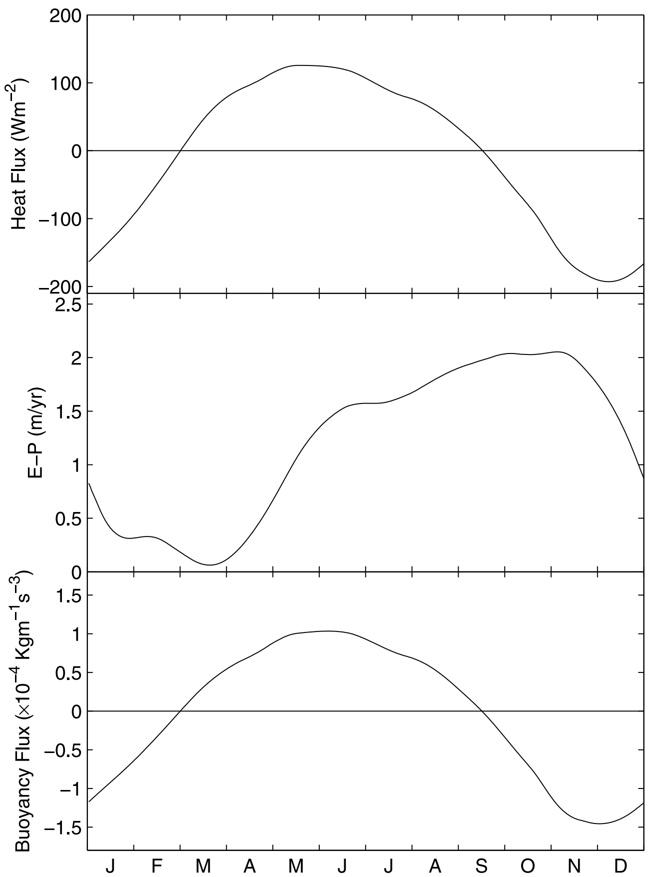

Figure 4. Daily surface fluxes averaged over the gulf from the 10th year model output in the climatological run.

gation to the strait are shown through seasonal vertical sections along the axis of the gulf and along the 54.6°E longitude.

[11] We will begin with a discussion of the axial sections from the climatological run driven by the corrected COADS forcing [see *Yao and Johns*, 2010]. The winter (maximum in February) cooling leads to a decrease of temperature and nearly isothermal condition in the northern gulf, and deep water is ventilated (Figure 5a). In the model, the coldest water with temperature of 19°C is formed in the northern end of the gulf, with salinity >42 psu and $\sigma_\theta > 30$. The salinity is also almost vertically uniform in the northern gulf, but exhibits a vertical gradient after 600 km, near 52°E. The vertical density stratification in the eastern gulf during winter is controlled primarily by the salinity. The salty water released from the southern banks is manifested as a bottom salinity maximum located between 700 and 800 km on the axial section. The winter conditions in the gulf resemble the solution of *Phillips*' [1966] two-dimensional convective overturning circulation model. The strong buoyancy loss (Figure 4) maintains a strong convective mixing field down to the bottom in the northern gulf. The general structure of the temperature and salinity fields agree well with the winter axial section observed by *Reynolds* [1993] (Figure 1a), but the model temperature is about 3°C warmer and the salinity is about 1 psu higher than the respective values of *Reynolds*' [1993] winter surveys.

[12] In spring (May) the seasonal thermocline in the gulf starts to build up and the export of the cold and salty bottom waters along the trough is obvious (Figure 5b). In summer (August) the surface temperature in the gulf exceeds 35°C and is warmer than the surface water in the Gulf of Oman (Figure 5c). The seasonal thermocline is fully established by this time and is located at a depth around 40 m. The deep waters are generally warmer than their winter and spring values and are totally isolated from the surface. The modified IOSW moves almost to the northern end of the gulf, and the salinity front along the axis is disturbed by eddies with a saltier core. The strong stratification in summer leads to weak vertical buoyancy flux and no deep water is formed at this time. Meanwhile, the cold, dense water formed in the winter continues to propagate toward the strait.

[13] In fall (November; note that there are no historical observations during fall), the surface heat loss leads to the deepening of the turbulent mixed layer (Figure 5d). While the surface temperature is decreased, the bottom water temperature is increased by continued downward turbulent heat flux. The bottom density is significantly reduced and the density gradient between the gulf and the Gulf of Oman reaches its lowest value of the year. Relatively warm water now appears in the descending outflow through the strait. This leads to a sharp vertical temperature front in the strait around 40 m, a similar feature to that seen in the fall observation in the strait region by *Pous et al.* [2004, Figure 11].

[14] The formation and the export of the dense water from the southern banks is shown in the winter cross section along 54.6°E in Figure 6a. The temperatures on the shelf are almost vertically uniform, but a trace of slightly warmer deep water formed during the fall is found in the trough. The lowest temperature is about 22°C and is warmer by about 3°C than the deep winter water in the northern gulf. Both the isohalines and isopycnals tilt down from the shelf forward the trough, consistent with the bottom flow from the southern banks as shown later in Figure 7. As in the northern gulf, the temperature is about 3°C warmer than that of *Reynolds*' [1993] observation and the salinities are slightly higher on average by about 0.5 psu.

[15] In summer (Figure 6b), the waters on the southern banks become very warm and the connection with the deep water is cut off. At this time the deep water entering the axial trough is from the northern gulf only. The warm yet salty water on the southern banks mixes laterally with the modified IOSW, which now moves much farther onto the banks in summer [*Yao and Johns*, 2010, Figure 10].

[16] The bottom salinity and current fields (Figure 7) depict the seasonal export pathways of the dense waters. In winter, dense waters are formed both in the northern gulf and southern shallow banks, as indicated by the sinking current around the Arabian coast west of Qatar and along the section along approximately 54.6°E, respectively. Both branches of sinking currents reach the axial deep trough and are clearly evident by their high-salinity signal. The waters from the southern gulf have slightly higher salinity as a result of the shallowness of the region and less mean exchange rate with the fresher waters in the interior of the gulf as shown later in Figure 11. Being closer to strait, the salt tongue released from the southern banks reaches the strait after about one month of advection. Animations of the bottom flow and salinity fields

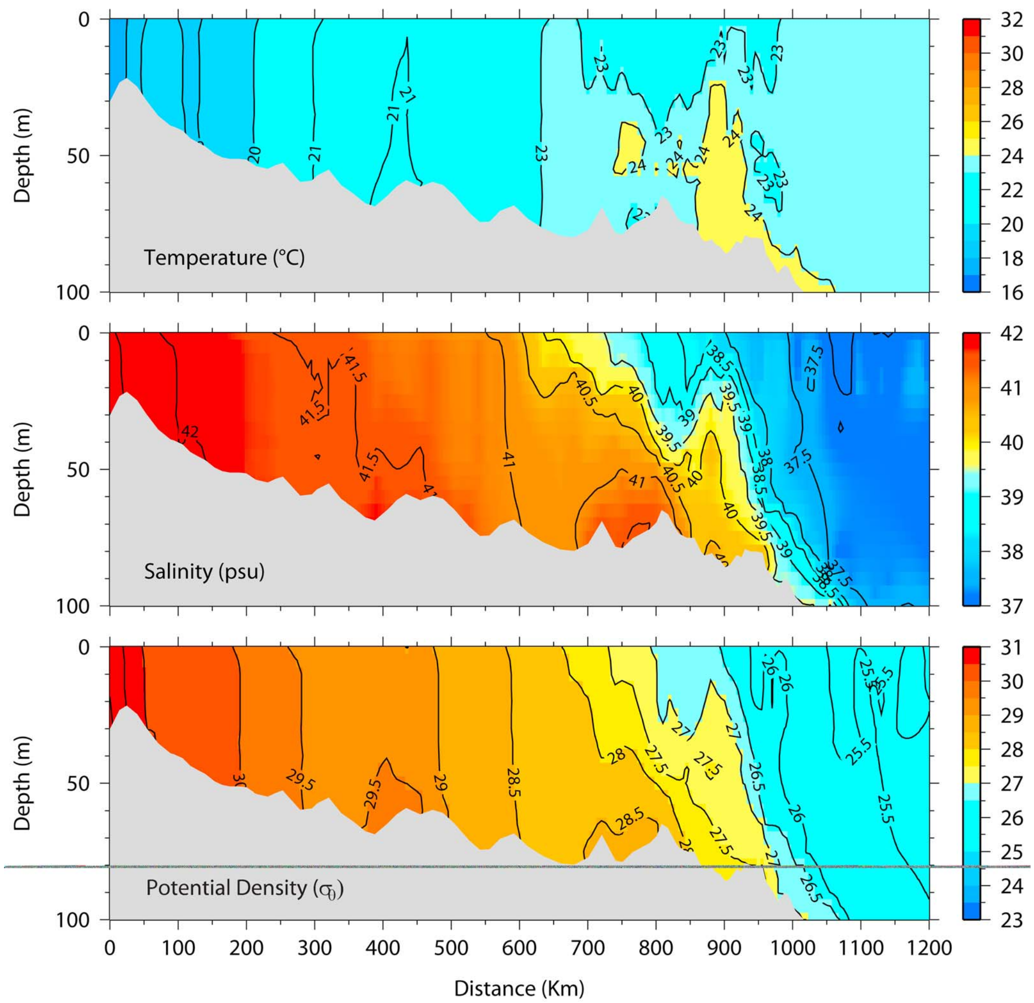

Figure 5a. Vertical axial sections for the middle of February in the 10th year model output in the climatological run. The strait is located around the distance of 900 km.

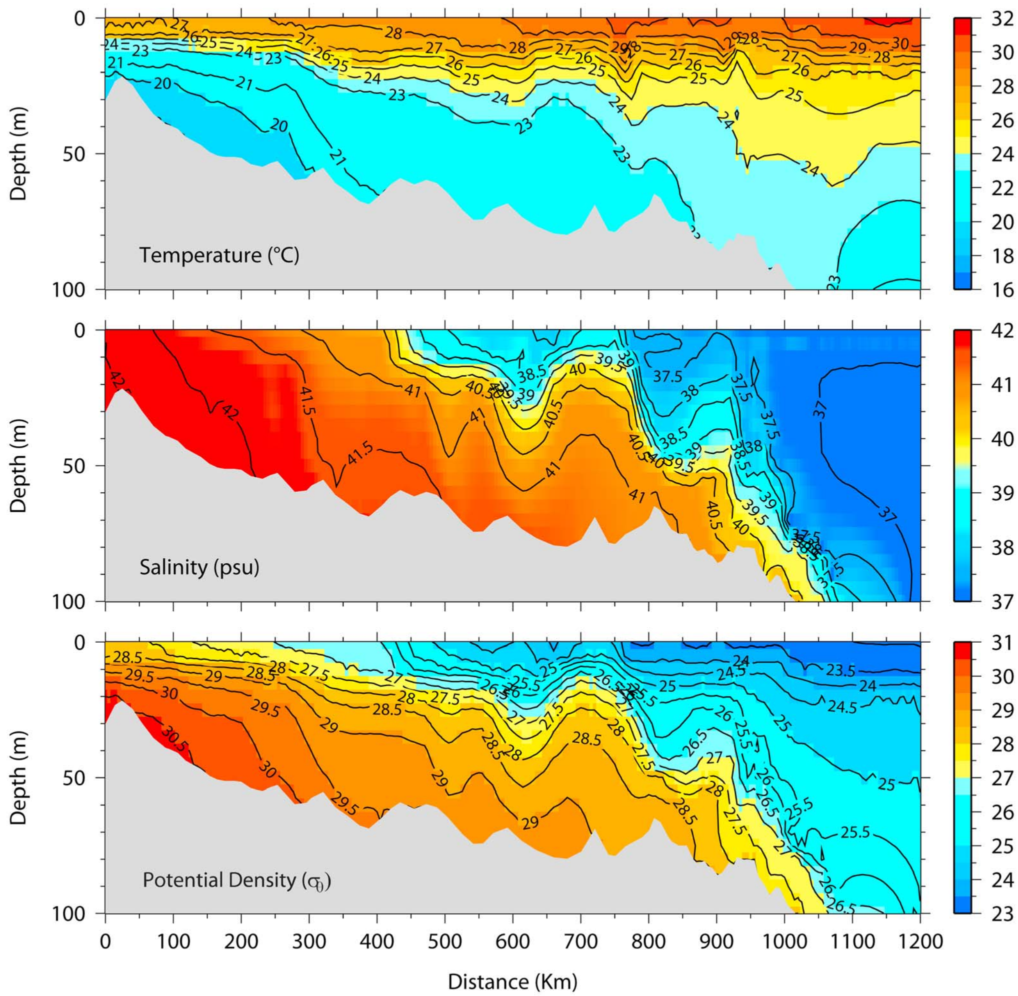

Figure 5b. Same as in Figure 5a but for the middle of May.

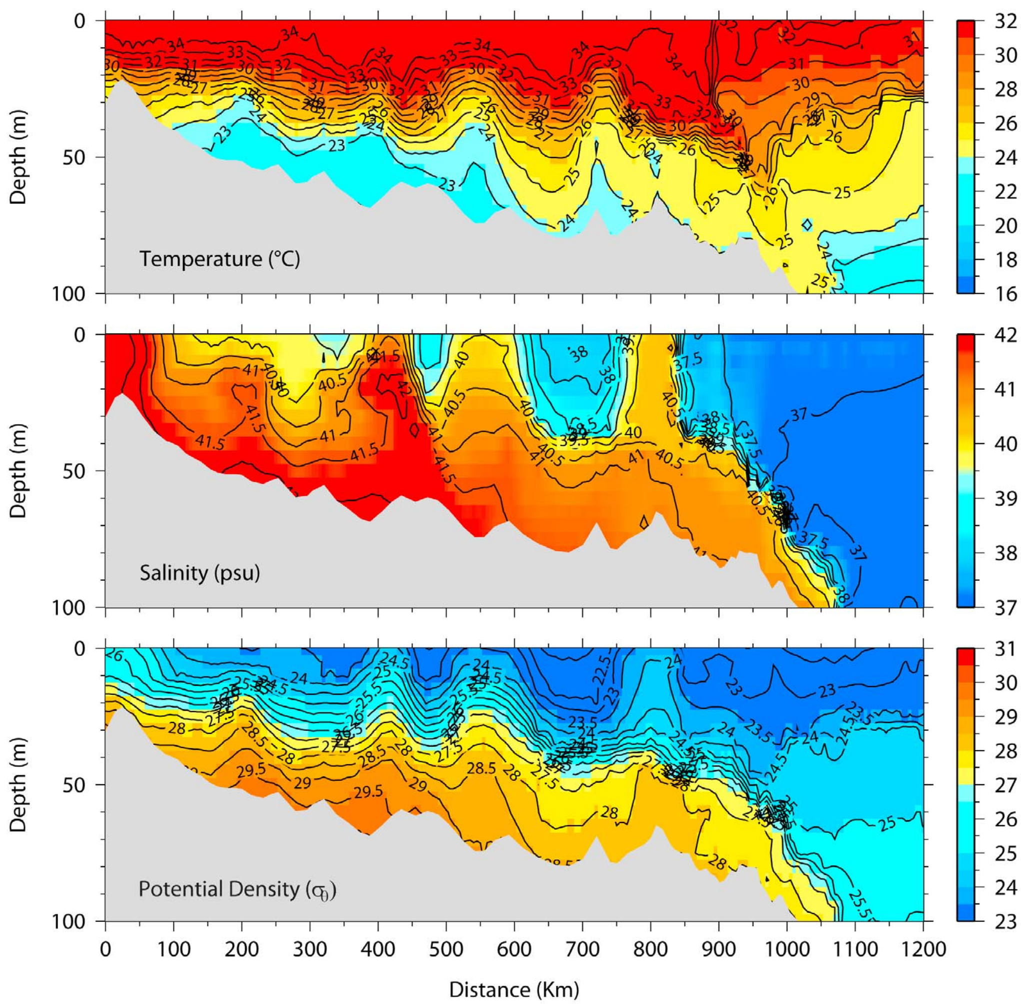

Figure 5c. Same as in Figure 5a but for the middle of August.

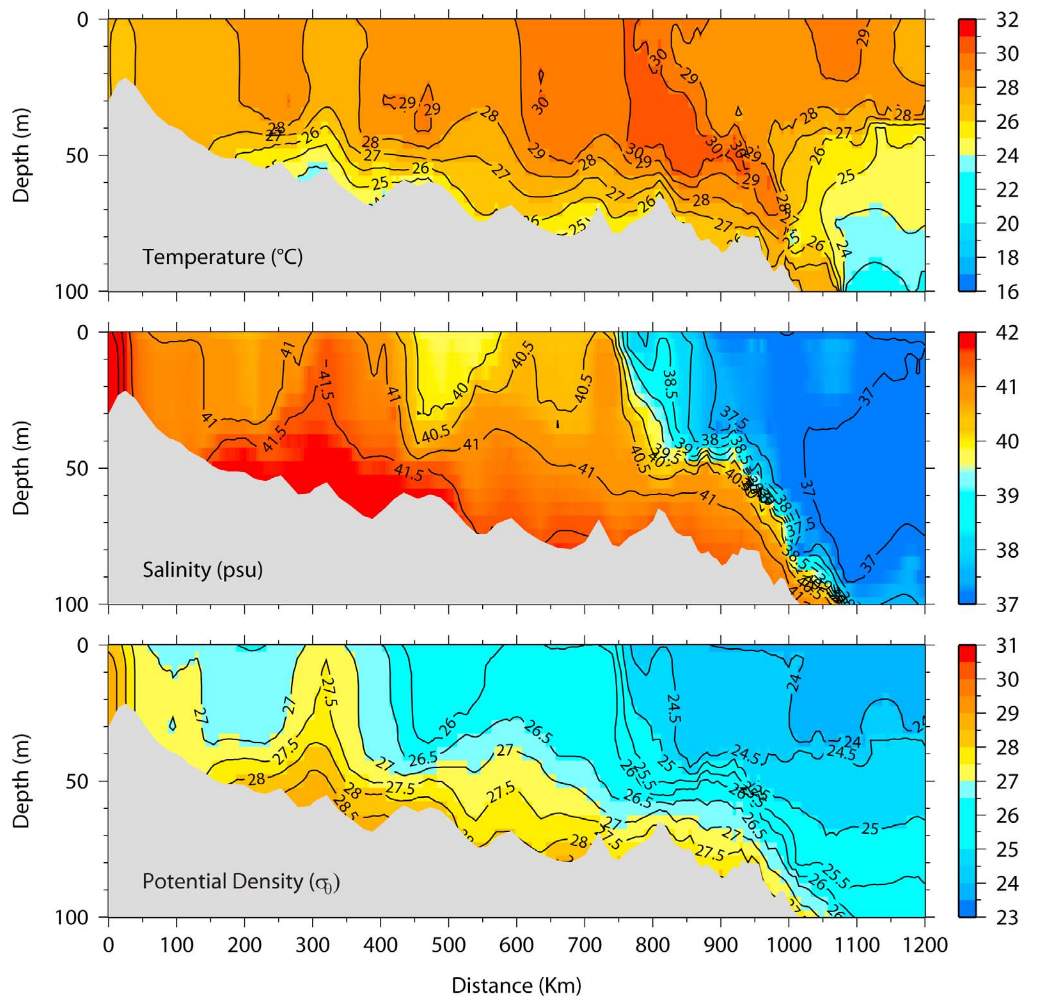

Figure 5d. Same as in Figure 5a but for the middle of November.

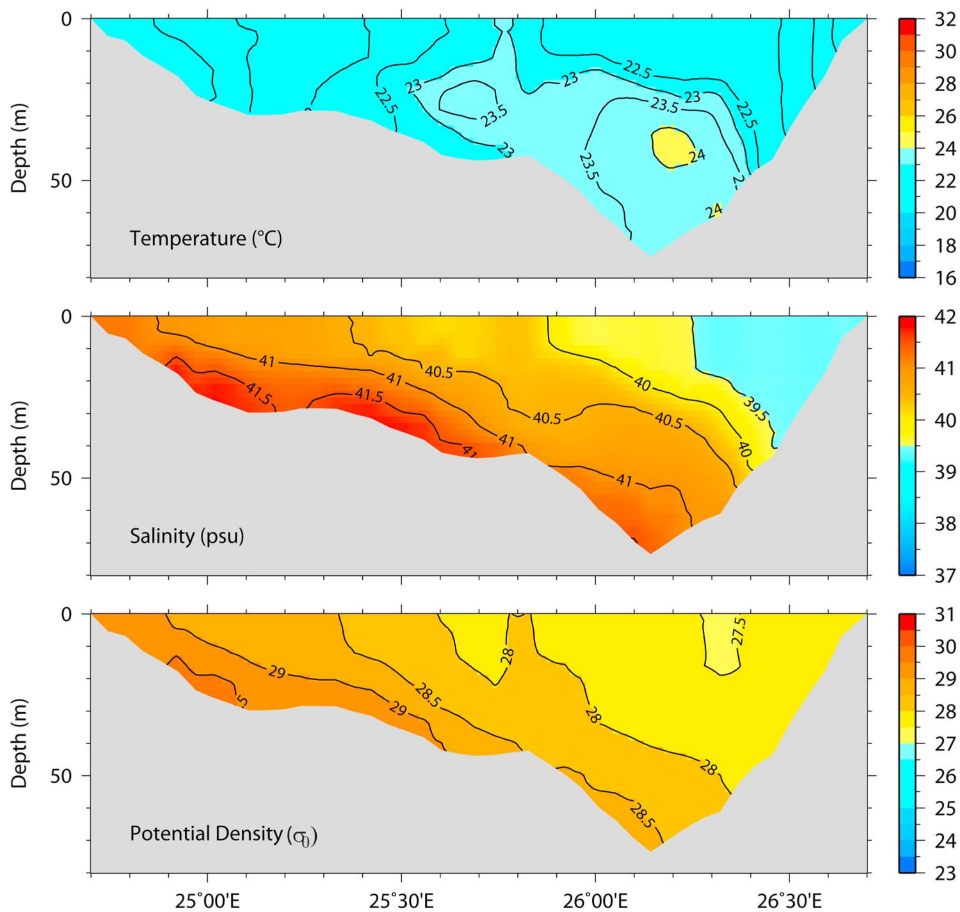

Figure 6a. Vertical sections along 54.6°E in the middle of February from the 10th year model output in the climatological run.

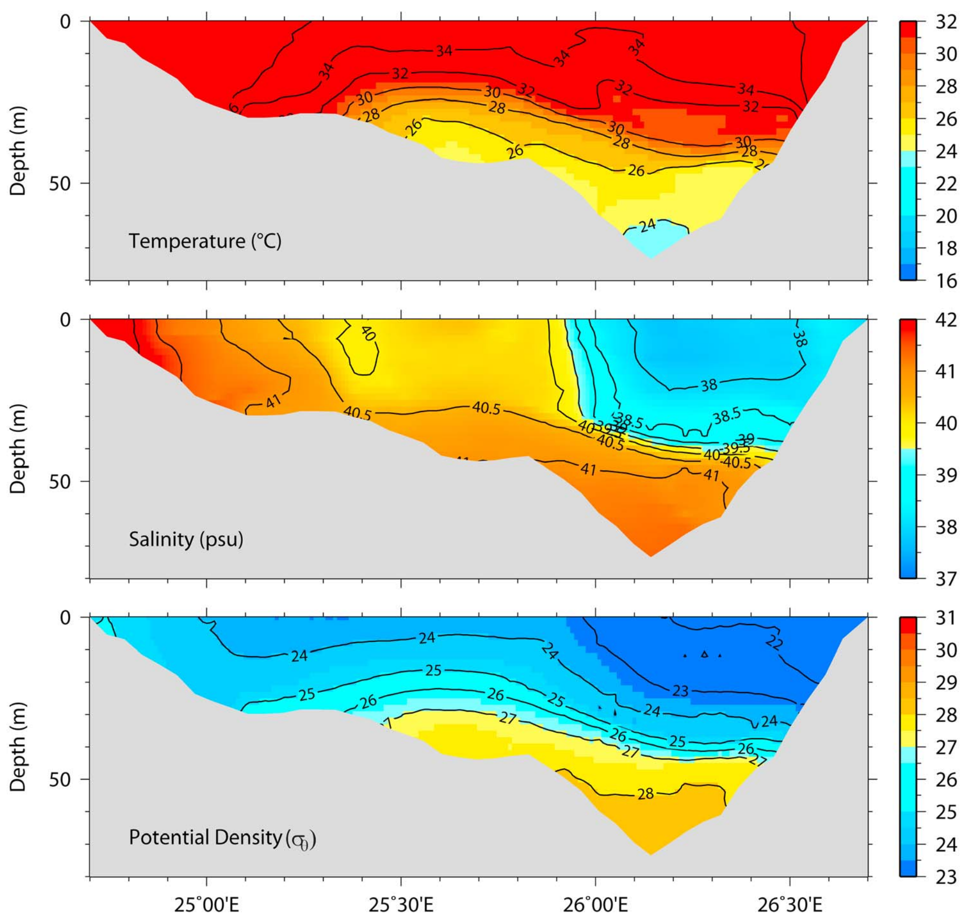

Figure 6b. Same as in Figure 6a but for the middle of August.

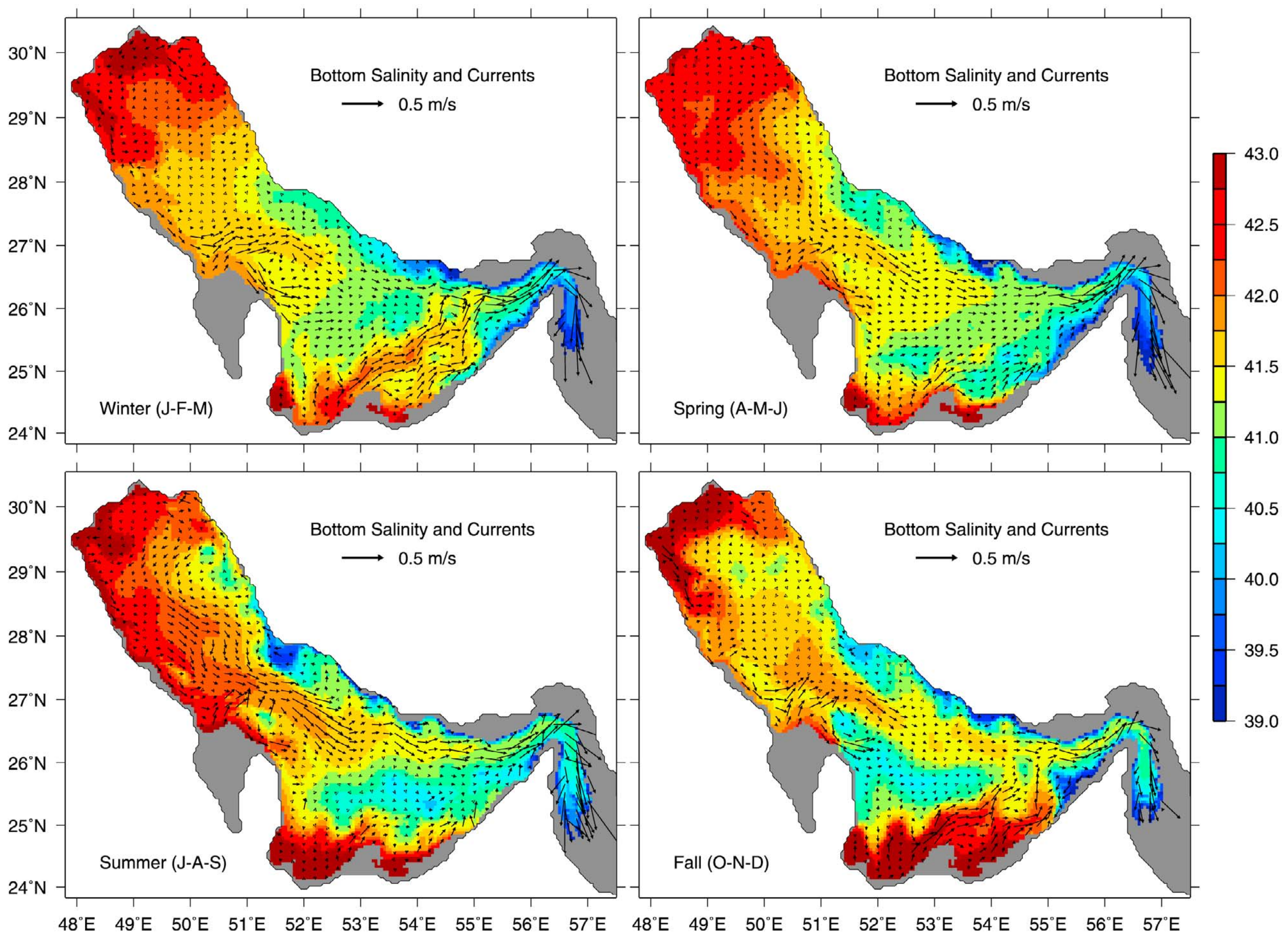

Figure 7. Seasonally averaged bottom salinity fields and currents from the 10th year model output. Area with salinity <39 psu is marked as grey.

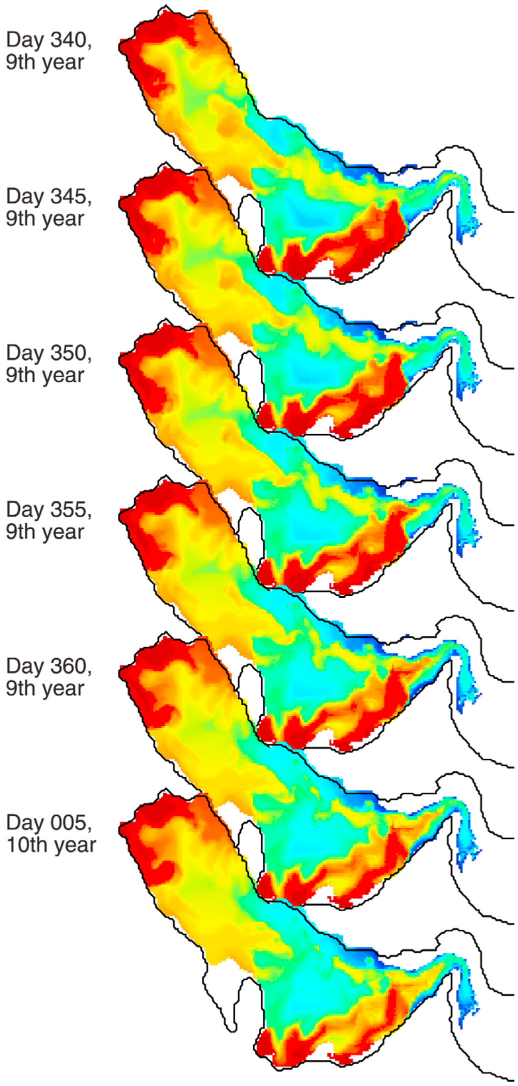

Figure 8. A series of snapshots of the bottom salinity showing the spillage of high-salinity waters from the southern banks to the strait during late fall and winter.

(Figure 8) show that the spilling of dense water off the southern banks is episodic, causing high-salinity pulses in the strait during winter, strikingly similar to the observations of *Johns et al.* [2003]. Before reaching the axial trough, the salty bottom outflow from the southern banks is concentrated in a region between 54 and 55.4°E, where a local trench connects the southern banks to the deep axial trough in the gulf (Figure 3). In contrast, the high-salinity water from the northern gulf propagates along the axial trough toward the strait and takes considerably longer (~4 months) to reach the strait. The salty water from the northern gulf experiences stronger mixing as it travels along the axial trough to the strait and is manifested as less salty water in the strait. While the dense water associated with the salt tongue from the northern gulf continues to propagate in spring and summer, the sinking of dense water from the southern bank only occurs in winter. The high-salinity water at the southern banks becomes too light to sink due to spring and summer warming. Instead it is horizontally expanded as a subsurface recirculation that can reach the strait as indicated by the warm and saltier water along the upper southern side of the strait (See Figure 12b). The export of the dense waters in the bottom layers through the strait will be discussed in section 4.

3. Overturning Circulation in Density Classes

[17] In order to quantify the two branches of the overturning circulation, the two formation zones of dense water are bounded by two sections as in shown Figure 3. The exchanges through the two sections are complex, so the *net* volume transports across the sections are decomposed into density classes ranging from 23 to 29.5 with 0.5 unit interval, to illustrate the overturning circulation across the two sections. The mean properties of salinity and temperature for each density classes are also calculated.

[18] The monthly mean overturning circulation in density classes in February and August in the climatological run are shown in Figures 7 and 8. For the northern branch, the winter overturning circulation occupies a narrow range of densities. The inflow is in the range σ_θ = 29–29.5, and the outflow is in σ_θ = 29.5–30 with mean temperature of 20°C and salinity of 41.4 psu. The summer overturning circulation is more expanded in density space (Figure 9b). The inflow is at densities of σ_θ = 22.5–24.5 with salinity 39–40 psu and temperature 34–35°C. The deep outflow occupies the density space between σ_θ = 26.5–30 and is characterized by linearly increased temperature ranging from 22°C to 26.6°C and a nearly uniform salinity of 41.7 psu. In between the two density classes exists a surface outflow with σ_θ = 24.5–27 characterized by the high temperature and high salinity, which represents the surface high-salinity outflow along the Arabian coast west of Qatar [see *Yao and Johns*, 2010, Figure 10].

[19] The inflow onto the southern banks in winter occurs in a density range σ_θ = 27–28.5 with almost uniform temperature of 22.3°C and salinity ranging from 39.4 to 40.5 psu. The outflow occurs in density classes σ_θ = 28.5–29.5 with slightly colder temperature and higher salinity. The inflow above σ_θ = 29.5 may be associated with the water exchange with the gulf deep water in the deep part of the section. The overturning circulation in summer is concentrated in much lighter density space; the overturning cell is much weaker and mainly occurs between σ_θ = 23.5–24.5. The outflow is slightly colder and saltier than the inflow (Figure 10b).

[20] The overturning rate is defined as the sum of the total outflow of the dense waters in both branches. The annual cycles of the overturning rates for each of the two branches are plotted in Figure 11. The overturning circulation in both branches shows a significant seasonal variation. While the southern branch reaches its highest level of 0.08 Sv in February, the northern branch reaches it highest value of 0.12 Sv in August. The maximum overturning in the southern branch occurs about 3 months after the time of maximum buoyancy loss in the gulf (November–December; Figure 4), and coincides with the transition from net buoyancy loss to net buoyancy gain over the gulf (that is, when the accumulated buoyancy loss over the winter season reaches its maximum). However, the overturning in the northern branch substantially lags (by nearly 6 months) the end of the winter

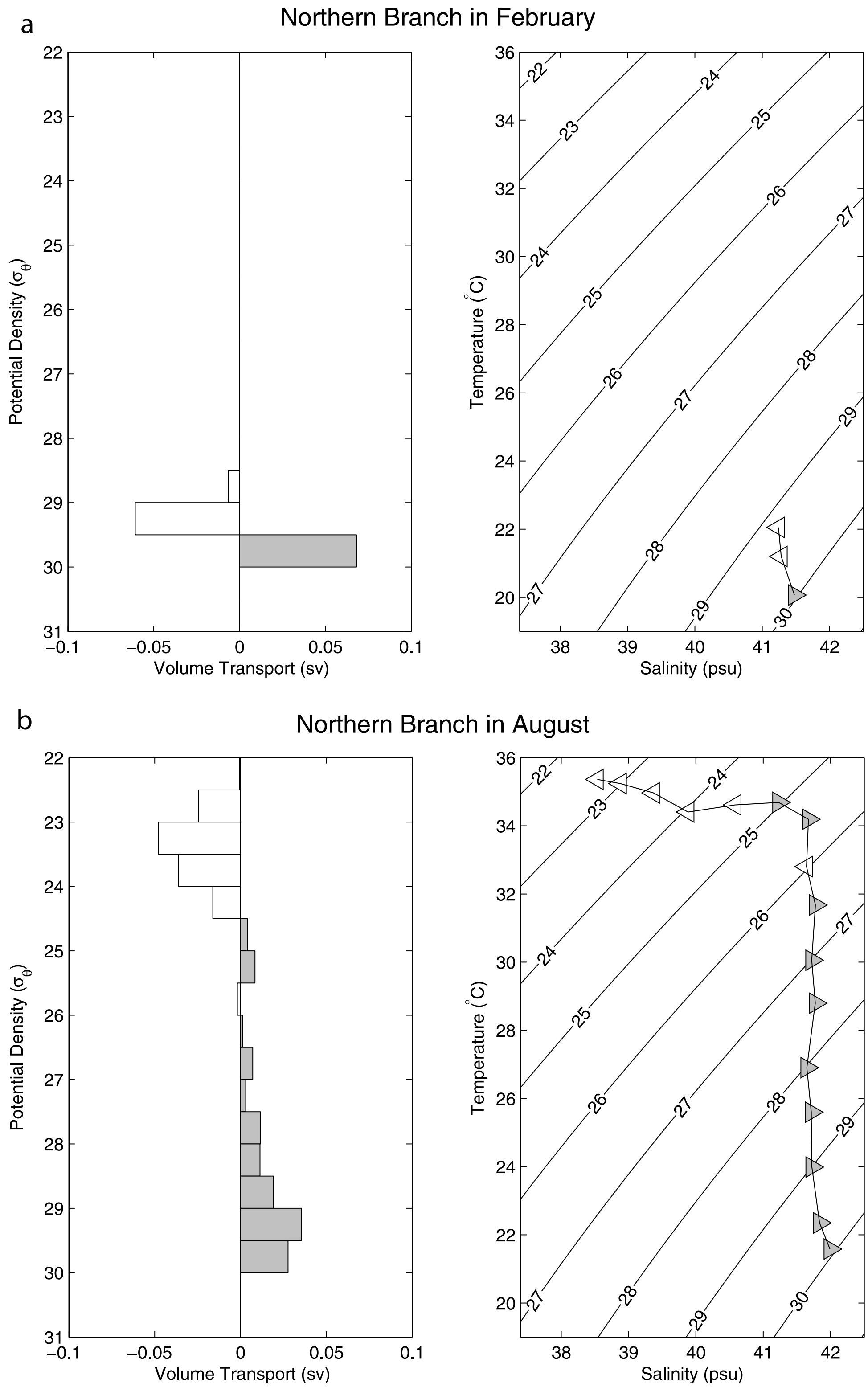

Figure 9. (a) Overturning circulation in density classes for the northern branch in February. (left) The integrated volume exchanges across the section shown in Figure 3 in the density classes. (right) The corresponding mean temperature and salinity for the transport are shown in the T-S diagram. The grey color represents the outflow. (b) Same as in Figure 9a but for August.

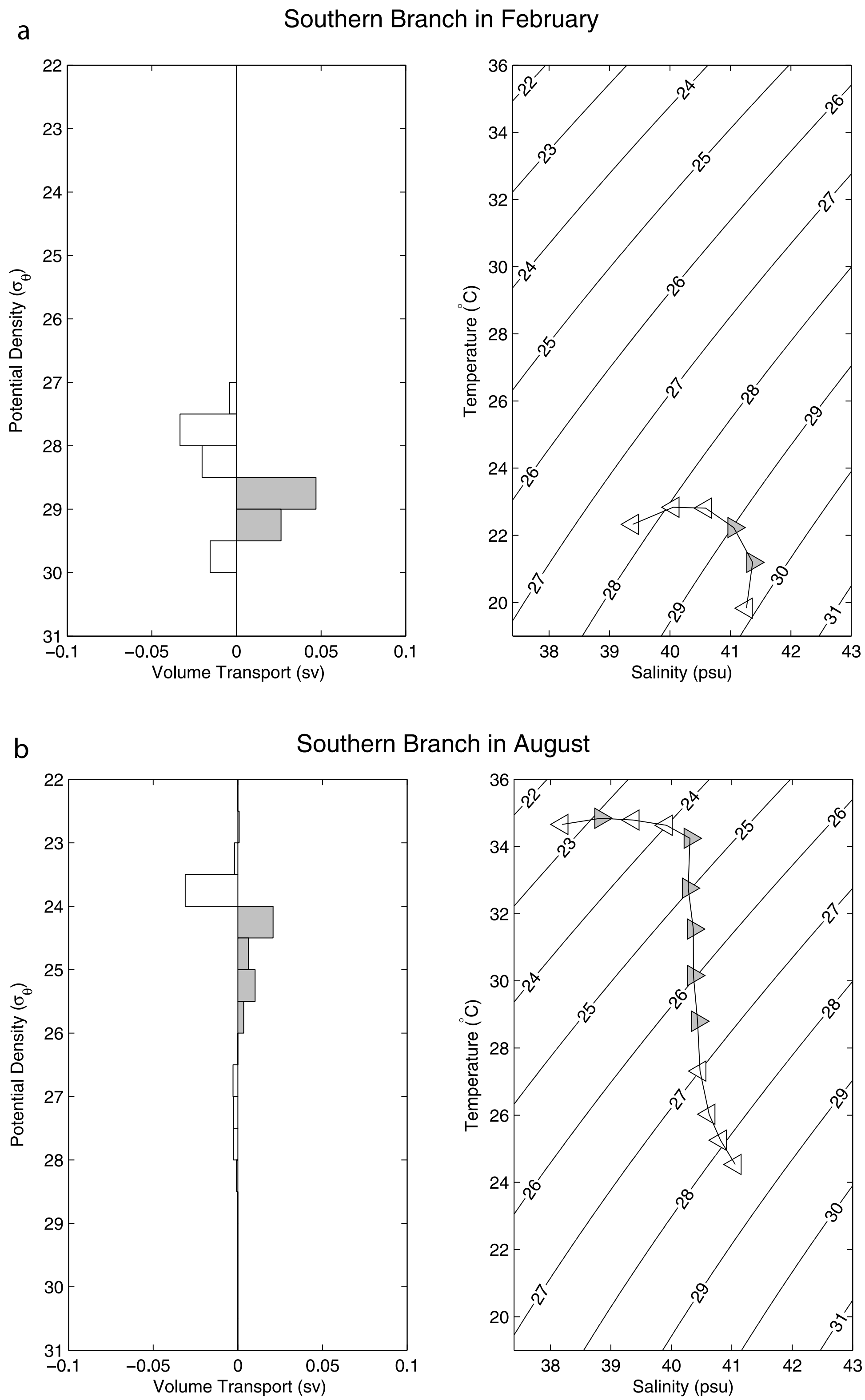

Figure 10. (a) Overturning circulation in density classes for the southern branch in February. (b) Same as in Figure 10a but for February.

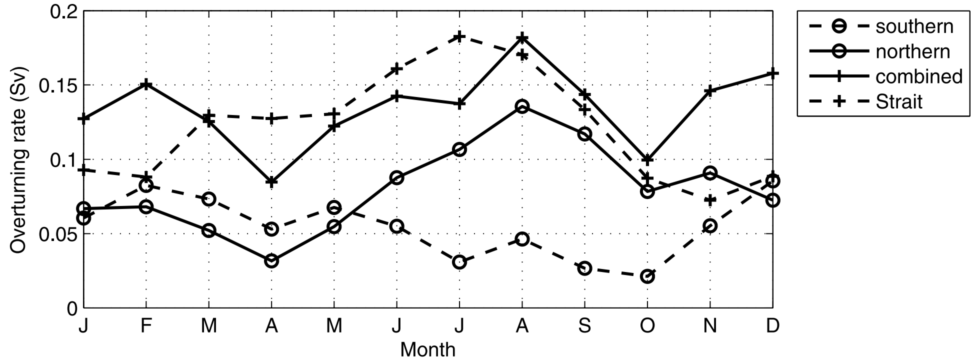

Figure 11. Annual cycle of the overturning rates in the density classes for the northern and southern branches and through the strait.

buoyancy loss season, due to the time it takes the dense waters to propagate from the northern gulf. The combined annual mean overturning rate of both branches, of 0.12 Sv, is the same as the model-derived deep outflow from the strait, as described in section 4.

4. Outflow in the Strait

[21] The water exchange through the strait is primarily driven by the density difference between the gulf deep water and the water in the Gulf of Oman. Therefore, the intensity of the exchange, and the water properties of the exchange flow, are determined by the water masses formed in the gulf. The strait is also the optimal location to compare the model and observational results. The exchange processes in the strait obtained from the simulation are presented in a similar fashion as in the observational study by *Johns et al.* [2003]. The monthly averaged sections of temperature, salinity, and velocity for February and August from the climatological run along 56°E across the strait are shown in Figures 12a and 12b, respectively.

[22] The water exchange structure generally compares favorably with the the observations available in the southern half of the strait [*Johns et al.*, 2003]. The temperature section in winter (February) is almost uniform and shows vertical inversion of warmer but saltier deep water. The salinity section shows a tilted across-strait structure with the deep salty outflow water banked against the southern side of the strait due to geostrophy. The outflow speed in winter is about 0.1 m/s, which is about a factor of 2 less than seen in the observation, for comparable monthly mean estimates. In the northern part of the strait, surface intensified inflow enters the gulf with maximum speed of also about 0.1 m/s.

[23] In summer (August) sections the temperature is well stratified (Figure 12b). The outflow is characterized by water with temperature < 27°C and salinity >40.5 psu. The outflow has a well defined core at depth of 50 m with maximum velocity of 0.2 m/s. The IOSW occupies the northern upper part of the strait and consists a very warm surface layer with temperature > 32°C and cooler subsurface layer with temperature around 29°C. The warm and salty water in the upper southern part of the strait is associated with the recirculation driven by the water from the southern banks.

[24] The time series of vertical profiles of temperature and salinity at the closest location to the *Johns et al.* [2003] mooring site are presented in Figure 13. The overall seasonal pattern agrees with the observation [see *Johns et al.*, 2003, Figure 7], but the temperature is about 2°C warmer than that in the observations both for the surface inflow and deep outflow. The deep cold water only persists to September, while it is present until November in the observations. This may be caused by excessive vertical penetration of the surface mixed layer under the strong buoyancy loss conditions in late fall (Figure 5d). The warm water in October and November with temperature about 31°C located at depths between 40 and 50 m is associated with the discharge of the warm and salty water from the southern banks, and it is also clearly present in the observations. The winter salinity pulses present in the observations [*Johns et al.*, 2003, Figure 7] are reproduced in the simulation (e.g., the January and March events in Figure 13). As can be seen from the bottom current in Figures 7 and 8, the salinity pulses are fed by the release of the salty water from the southern banks in winter. In contrast, a recent numerical study by *Thoppil and Hogan* [2009] found that short-term (several days to a few weeks) high-salinity outflow pulses occurs throughout the year, that are generated by the mesoscale cyclone eddies induced by fluctuations in the wind stress forcing. However, the high-salinity pulses observed by *Johns et al.* [2003] that have longer time scales of more than one month and only occur in winter in the observation, are suggested by our model to be produced by another mechanism, i.e., the winter discharge of the dense waters from the southern banks.

[25] The volume transports of the deep outflow across the section are integrated from 45 m to the bottom (Figure 14). Contrary to the fairly steady volume transport estimated from the observations, the volume transport of the simulated outflow in the climatological run exhibits significant seasonal variability. The transport is larger in summer than in fall and winter, and the maximum transport occurs in July with value of 0.15 Sv and the minimum transport of 0.07 Sv in January. The annual mean volume transport of the outflow is 0.12 Sv and is slightly weaker than the estimate of 0.15 Sv by *Johns et al.* [2003].

[26] The volume transports in the buoyancy-only forcing experiment and high-frequency experiment described by *Yao and Johns* [2010] follow a similar seasonal cycle as in the climatological run. The outflow in the high-frequency run has an annual mean value of 0.12 Sv, the same as the climatological run, but has weaker transport in the winter months

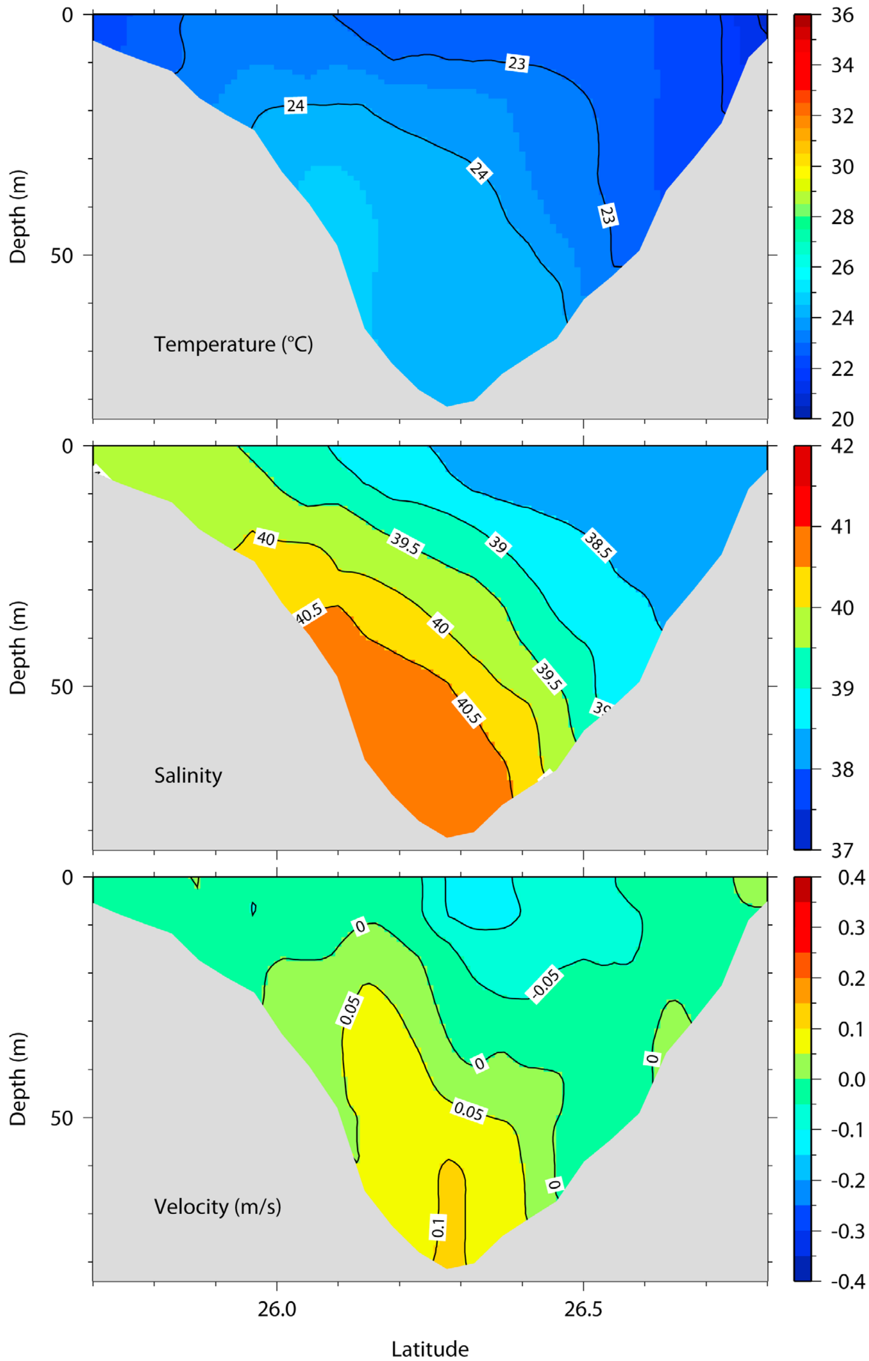

Figure 12a. Vertical sections across the strait (positions as shown in Figure 3) in February from the 10th year model output in the climatological run.

(December, January, and February) due to stronger winter vertical mixing. In density classes (Figure 11), the combined net overturning in the northern and southern gulf is equal to the export through the strait in the annual mean, but they show some differences in their seasonal cycle. In particular, the net overturning in winter (November–February) exceeds the deep water export from the strait, while during spring and summer the reverse is true. This is due to a phase lag in the export of the deep water produced in the overturning regions, as this water collects in the deep trough and moves to the strait. The net effect of this lag is to produce a stronger annual cycle in the strait outflow than in the overturning regions themselves, as deep water that is formed in winter is temporarily stored in the deep trough and continues to outflow in the spring and early summer.

[27] Dynamically, the forcing for the deep outflow through the strait depends on the pressure difference between the gulf and the Gulf of Oman at comparable depth. As seen in Figure 5d, the along strait density difference is weakest in November, consistent with the minimum deep outflow from

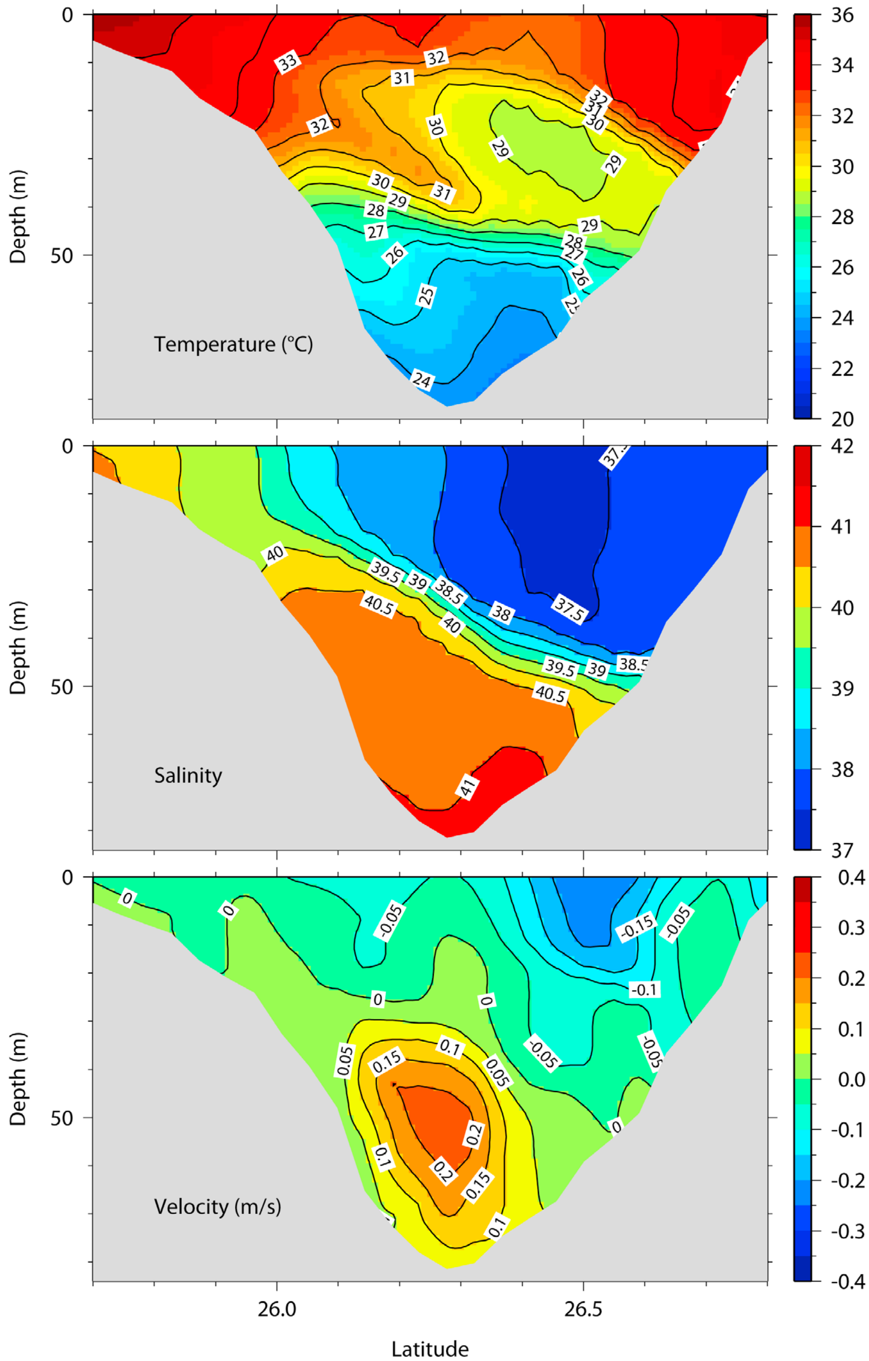

Figure 12b. Same as in Figure 12a but for August.

the strait. The pressure difference between the inside and outside of the gulf diagnosed from the climatological run is given in Figure 15. The inflow in the upper layer is associated with the negative pressure gradient from the Gulf of Oman to the gulf and there is a reversal of the pressure difference at 60 m, driving the deep outflow from the gulf. There is a seasonal pattern for the pressure difference at 60 m (Figure 15), with maximum difference in summer and minimum in winter. The seasonal bottom outflow in the strait (Figure 14) closely mirrors this cycle. Another possible mechanism that could contribute to the weaker outflow in fall and winter is friction associated with increased vertical eddy viscosity, where downward momentum flux from the upper layer caused by the vertical mixing weakens the baroclinic circulation [*Linden and Simpson*, 1988]. Both of these effects are linked to the strong surface heat loss in fall and winter and the enhanced vertical mixing it causes. The lack of any clear seasonal cycle in the observed outflows leaves the model results in doubt. On the other hand, the outflow deduced by *Johns et al.* [2003] is based on observations with duration of less than 1 year, from a single mooring with a significant gap in summer. Therefore longer observations in the strait are

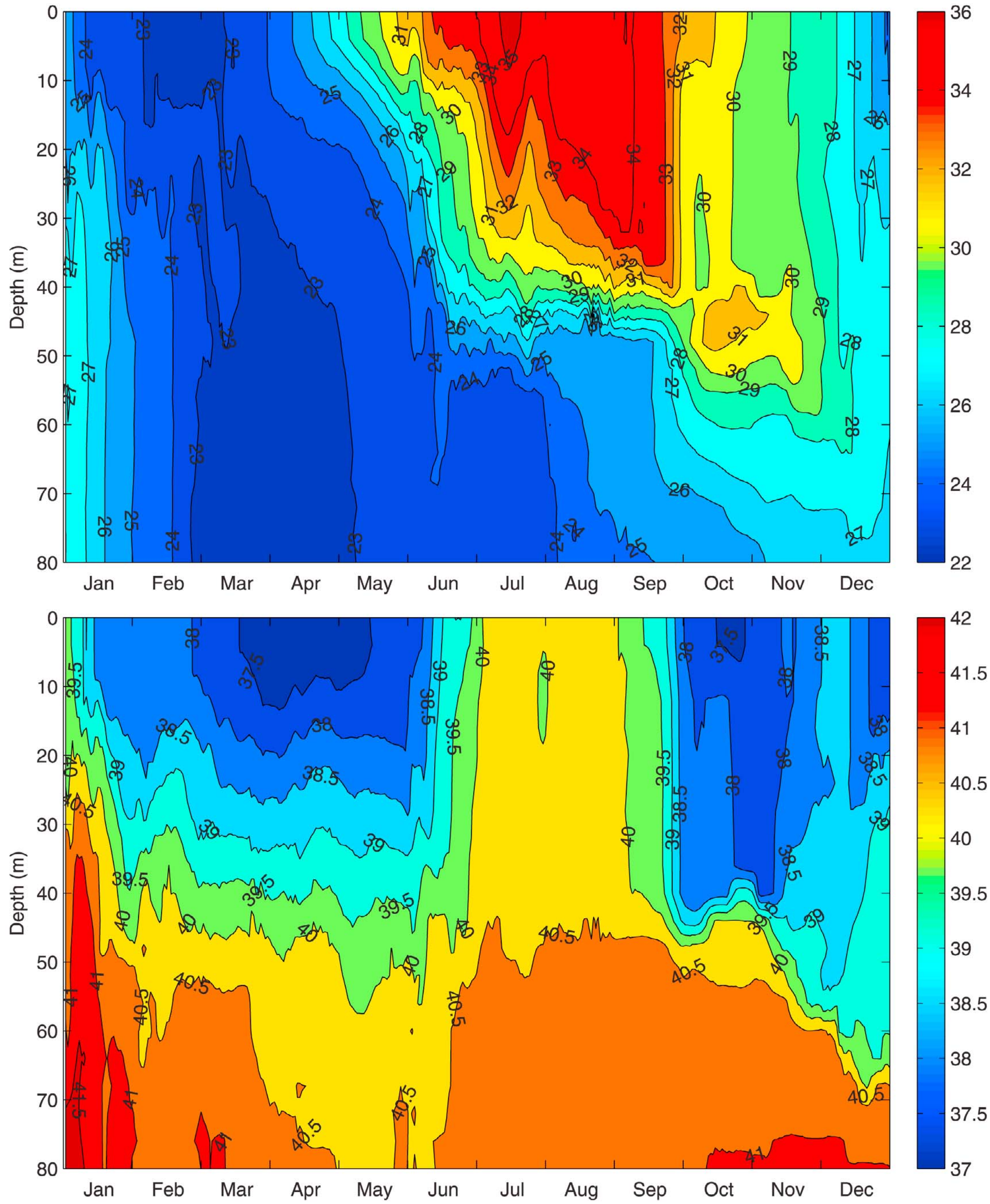

Figure 13. Time series of the temperature and salinity from the 10th year in the climatological run in the strait closest to the *Johns et al.* [2003] mooring site.

probably needed before the seasonal variability of the deep outflow can be conclusively established.

5. Discussions and Conclusions

[28] The gulf shares many aspects with other semienclosed, marginal seas, such as the Red Sea and the Mediterranean Sea, in that the general circulation is a coupled result of the wind-driven and thermohaline-driven flow, complex mesoscale eddy fields, and constricted water exchange with the open ocean. As suggested by the observational results of *Reynolds* [1993] and *Johns et al.* [2003], the circulation in the gulf exhibits unique features compared to its counterparts because of a much shallower basin and less constricted connection to the open ocean due to the wider strait and the absence of a significant sill. The main goal of this study is to provide a systematic seasonal picture of the circulation and water mass formation in the gulf and bridge the gap between the observed features in the gulf and in the strait. As the atmospheric forcing plays a critical role in driving the

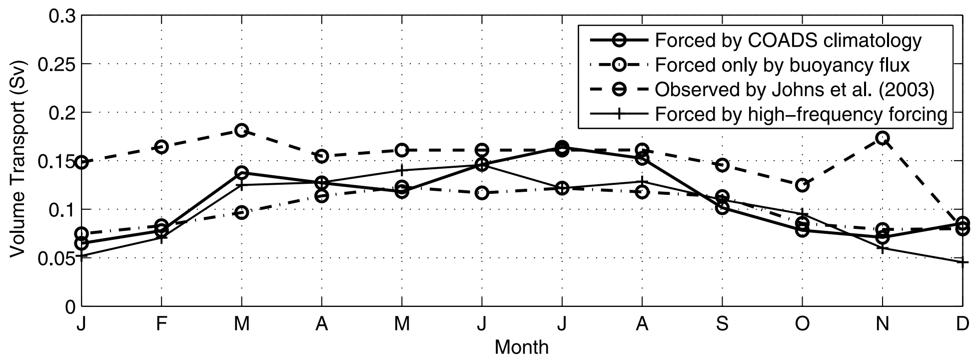

Figure 14. Volume transport of outflow below 45 m through the strait across 56°E from the models and the observations.

seasonal circulation and forming the water masses in the gulf, sensitivity experiments with different forcing are conducted to investigate the effect of the buoyancy flux, surface wind stress, and high-frequency fluctuations on the circulation.

[29] The surface current fields in the gulf are mainly driven by the low-salinity inflow of Indian Ocean Surface Water (IOSW) and are modified by the surface wind stress. The IOSW propagates in two branches into the gulf: (1) one along the Iranian coast toward the northern gulf and (2) one veered onto the southern banks. As suggested by the sensitivity experiment forced only with buoyancy flux of *Yao and Johns* [2010], this veering is driven by Ekman drift due to the prevailing northwesterly winds. These branches of inflow form two cyclonic gyres in the northern gulf and in the southern gulf. The surface inflow exhibits significant seasonal variability. The inflow is more intensified along the Iranian coast in summer than in winter, suggesting more transport of IOSW to the northern gulf. The inflow is shifted more toward the southern gulf in winter due to the stronger northwesterly winds.

[30] Corresponding to the two branches of the inflow, the IOSW is transformed into hypersaline waters with salinity >41 psu by the fresh water loss in the northern end of the gulf and on the southern shallow shelf. The seasonal cycle of the heat fluxes determines the seasonal variation of the density of these two sources of hypersaline water and their export as outflows toward the strait. In winter, cold and saline waters are formed both in the northern gulf and in the southern gulf. The dense waters from the southern gulf with higher-salinity spill into the trough as a sporadic bottom outflow in winter that leads to the salinity pulses in the strait as observed. While the dense waters formed in the northern gulf propagate toward the strait along the trough through the whole year, the waters in the southern banks are so warmed up in summer that they are exported as a warm yet salty intermediate depth water in late summer and fall. For the two branches of overturning circulation, the northern branch peaks in August with strength of 0.13 Sv and the southern branch peaks in February and December with strength of 0.08 Sv.

[31] The water exchange through the strait is driven by the outflow formed from these dense water sources and compensating inflow from the Gulf of Oman. The structure in winter is basically a two-layer exchange with inflow of salinity <38.5 psu in the upper northern part of the strait and deep outflow with temperature about 24°C and salinity

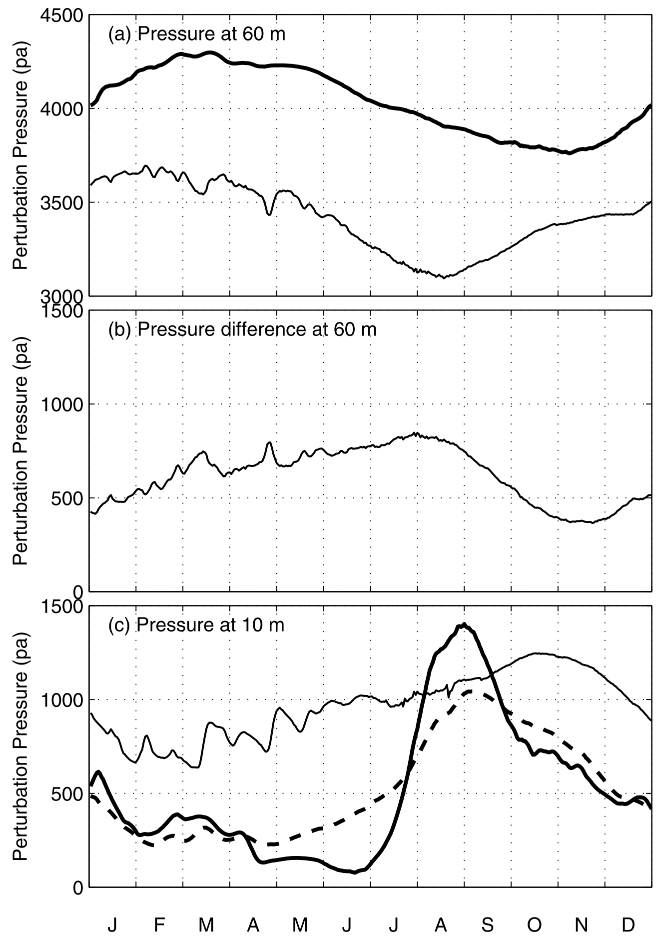

Figure 15. Annual cycle of perturbation pressure (with a reference to a static pressure with constant density) inside and outside the gulf. (a) Perturbation pressure at 60 m, with the thin line indicating pressure outside the gulf and the thick line indicating pressure inside the gulf. (b)Pressure difference at 60 m and (c) perturbation pressure at 10 m, with the thin lines indicating pressure outside the gulf, the thick lines indicated pressure inside the gulf, and the dashed line indicating the pressure averaged along 54.6°E to reduce the effect of surface eddies. See Figure 3 for the locations of the diagnosing positions.

40.5 psu banked against the deep southern side of the strait, consistent with observations. In summer and fall, in addition to the upper inflow of theIOSW and deep outflow, the warm and salty waters from the southern banks feed outflow in the intermediate depths between 10 and 40 m in the southern part of the strait, again consistent with observations. The variability of the exchange flow as shown in Figure 13 reflects the variation of the formation and export of the dense water formed in the gulf. The deep cold outflow with temperature < 26°C exists from January to November, and the salinity pulses in the salinity time series in winter occur due to discharge of the salty waters from the southern banks. The seasonal pattern of the water properties generally agree with the observations by *Johns et al.* [2003]. However, in the climatological run the temperature is about 3°C warmer than in the observational results and the salinity is 1 psu higher. The temperature and salinity fields agree better with the observations in the experiment driven by the high-frequency forcing [see *Yao and Johns*, 2010].

[32] The outflows integrated over the lower part of the strait from all three experiments show similar seasonal variability, with maximum transport in summer and minimum transport in winter, suggesting the water exchange through strait is dominantly driven by the buoyancy flux over the gulf. The monthly outflow transport is lower in winter and fall than the estimates by *Johns et al.* [2003], resulting a lower annual mean transport of 0.12 Sv. The annual cycle is considerably different from that of *Chao et al.* [1992], in which it is suggested the outflow peaks in March following the annual cycle of the evaporation. (Their annual cycle of evaporation is also considerably different from this study). However, these simulations suggest that the seasonal heat flux primarily controls the seasonality of the outflow. While during the winter cooling period, the overturning circulation is characterized by strong convection, the summer overturning circulation is driven by the subduction of the dense water formed in winter. The lower transport in fall and winter in the model is linked to a reduced pressure difference between the gulf and in the Gulf of Oman. Another possible reason is increased internal friction in winter between the inflow and outflow due to the increased vertical mixing. Both of the effects are related the heat loss in late fall and winter. On the other hand, the estimate by *Johns et al.* [2003] has a relatively large error bar due to the limited coverage ofthe outflow in the strait and the observation may not represent a typical year, so further measurements of long-term exchange through the strait are needed to firmly validate the numerical results.

[33] **Acknowledgments.** We would like to thank Alan Wallcraft, George Halliwell, and Eric Chassignet for the invaluable assistance on HYCOM. We also want to thank R. Michael Reynolds for the survey data in the gulf. We are extremely grateful to Wei Zhao and Shuyi Chen for providing the MM5 model output for the gulf region. This work was supported by grant N00014-0310319 from the Office of Naval Research.

References

Bower, A. S., H. D. Hunt, and J. F. Price (2000), Character and dynamics of the Red Sea and Persian Gulf outflows, *J. Geophys. Res.*, *105*, 6387–6414, doi:10.1029/1999JC900297.

Brewer, P. G., and D. Dyrssen (1985), Chemical oceanography of the Persian Gulf, *Prog. Oceanogr.*, *14*, 41–55.

Chao, S. Y., T. W. Kao, and K. R. Al-Hajri (1992), A numerical investigation of circulation in the Arabian Gulf, *J. Geophys. Res.*, *97*(C7), 11,219–11,236.

Johns, W. E., F. Yao, D. B. Olson, S. A. Josey, J. P. Grist, and D. A. Smeed (2003), Observations of seasonal exchange through the Straits of Hormuz and the inferred heat and freshwater budgets of the Persian Gulf, *J. Geophys. Res.*, *108*(C12), 3391, doi:10.1029/2003JC001881.

Kämpf, J., and M. Sadrinasab (2006), The circulation of the Persian Gulf: A numerical study, *Ocean Sci.*, *2*, 27–41.

Linden, P. F., and J. E. Simpson (1988), Modulated mixing and frontogenesis in shallow seas and estuaries, *Cont. Shelf Res.*, *8*(10), 1101–1127.

Phillips, O. M. (1966), On turbulent convection currents and the circulation of the Red Sea, *Deep Sea Res. Oceanogr. Abstr.*, *13*, 1149–1160.

Pous, S. P., X. Carton, and P. Lazure (2004), Hydrology and circulation in the Strait of Hormuz and the Gulf of Oman—Results from the GOGP99 Experiment: 1. Strait of Hormuz, *J. Geophys. Res.*, *109*, C12037, doi:10.1029/2003JC002145.

Prasad, T. G., M. Ikeda, and S. P. Kumar (2001), Seasonal spreading of the Persian Gulf Water mass in the Arabian Sea, *J. Geophys. Res.*, *106*, 17,059–17,073, doi:10.1029/2000JC000480.

Reynolds, R. M. (1993), Physical oceanography of the Gulf, Strait of Hormuz, and the Gulf of Oman—Results from the *Mt Mitchell* expedition, *Mar. Pollut. Bull.*, *27*, 35–59.

Rochford, D. J. (1964), Salinity maxima in the upper 1000 metres of the north Indian Ocean, *Aust. J. Mar. Freshwater Res.*, *15*, 1–24.

Swift, S. A., and A. S. Bower (2003), Formation and circulation of dense water in the Persian/Arabian Gulf, *J. Geophys. Res.*, *108*(C1), 3004, doi:10.1029/2002JC001360.

Thoppil, P. G., and P. J. Hogan (2009), On the mechanisms of episodic salinity outflow events in the Strait of Hormuz, *J. Phys. Oceangr.*, *39*, 1340–1360.

Wyrtki, K. (1973), Physical oceanography of the Indian Ocean, in *The Biology of the Indian Ocean*, *Ecol. Stud. Ser.*, vol. 3, edited by B. Zeitschel, pp. 18–36, Springer, Berlin.

Yao, F., and W. E. Johns (2010), A HYCOM modeling study of the Persian Gulf: 1. Model configurations and surface circulation, *J. Geophys. Res.*, *115*, C11017, doi:10.1029/2009JC005781.

W. E. Johns and F. Yao, Division of Meteorology and Physical Oceanography, Rosenstiel School of Marine and Atmospheric Science, University of Miami, 4600 Rickenbacker Cswy., Miami, FL 33149, USA. (bjohns@rsmas.miami.edu; fyao@rsmas.miami.edu)