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Formation and circulation of dense water in the Persian/Arabian Gulf

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[1] The nature and circulation of water masses in the Persian/Arabian Gulf (hereinafter referred to as the Gulf) is investigated by examination of a historic database of hydrographic observations. The densest water forms in winter at the northern end of the Gulf rather than along the warmer southern and western coasts. With the exception of small amounts of water directly above the seafloor, most water flowing out of the Gulf mixes across a density front that separates Gulf Deep Water within the Gulf from the Indian Ocean Surface Water (IOSW). Contrary to previous inferences, the seasonally variable incursion of IOSW into the Gulf peaks in late spring. This timing may be due to seasonal changes in sea surface slope driven by variations in evaporation rate. In order to explain mooring results published elsewhere that show relatively small seasonal changes in the volume flux through the Strait of Hormuz (hereinafter referred to as the Strait), we suggest that this flux is driven by the difference between the density of Gulf Deep Water in the interior of the basin and water at comparable depths outside the Gulf. This density difference varies less than 15% during the year. High rates of vertical mixing in the Strait extend about 200 km westward in response to topographic constriction of tidal flows by islands and shoals. *INDEX TERMS:* 4219 Oceanography: General: Continental shelf processes; 4223 Oceanography: General: Descriptive and regional oceanography; 4243 Oceanography: General: Marginal and semienclosed seas; 4283 Oceanography: General: Water masses; *KEYWORDS:* Persian/Arabian Gulf, Strait of Hormuz, water mass formation, density gradients, seasonal variability, vertical mixing

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1. Introduction

[2] High-salinity water flows out of the Persian/Arabian Gulf (hereafter called the Gulf) and spreads at 200–350 m depth within the Gulf of Oman in the northeastern Indian Ocean [*Rochford*, 1964; *Wyrtki*, 1971; *Qasim*, 1982; *Premchand et al.*, 1986; *Bower et al.*, 2000; *Prasad et al.*, 2001]. The injection of Gulf overflow water affects the stability of the Indian Ocean thermocline and introduces oxygen-rich water at a depth characterized farther east by extreme oxygen-depletion due to decay of surface layer primary production [*Wyrtki*, 1973; *Olson et al.*, 1993]. The density and thus the depth of Gulf overflow water is determined by its source characteristics as it flows through the Strait of Hormuz and by subsequent mixing processes on the continental shelf and slope off Oman and Iran [*Bower et al.*, 2000].

[3] The nature of the overflow water and dynamics of the exchange with the Indian Ocean depend on the contrast between processes affecting water properties in the shallow, land-locked Gulf basin and in the deeper, unconfined western Gulf of Oman. The Gulf, with an average depth of ~35 m and maximum depths of 110–160 m in current scoured channels near islands, is connected to the western Gulf of Oman through the Strait of Hormuz (hereafter, the Strait; Figure 1). Tectonic-driven subsidence deepened the seafloor locally in the Strait to 200–300 m and produced a 70–95 m deep trough along the Iranian side of the southern half of the Gulf (Figure 2) [*Ross et al.*, 1986]. The Strait and the basin trough are separated by a broad sill at about 86 m near 55°30′E in the western approaches to the Strait. A southward widening channel leads from the Strait south across a series of sills (water depth of ~110 m) and shallow basins to the shelf edge [*Seibold and Ulrich*, 1970]. Although water is deeper in the channel than in the Gulf, the narrow Strait of Hormuz restricts water exchange and isolates the Gulf from well-mixed water masses in the northern Indian Ocean.

[4] The climate in the Gulf region is arid, resulting in an excess of evaporation over precipitation plus river runoff. Although estimates of fresh water flux are quite variable, the following published values give a sense of the balance: evaporation rate 1.4 m/yr [*Privett*, 1959], river runoff 0.15–0.46 m/yr [*Hartmann et al.*, 1971; *Chao et al.*, 1992; *Reynolds*, 1993], and precipitation 0.07–0.1 m/yr [*Hartmann et al.*, 1971; *Reynolds*, 1993]. The relatively high evaporation combined with restricted exchange with the open ocean leads to formation of a saline, dense water mass (Gulf Deep Water) and a reverse estuary circulation through the Strait of Hormuz. Within the Strait, flow of dense water out of the Gulf is mostly confined to the southern side of the channel by geostrophy [*Emery*, 1956; *Chao et al.*, 1992],

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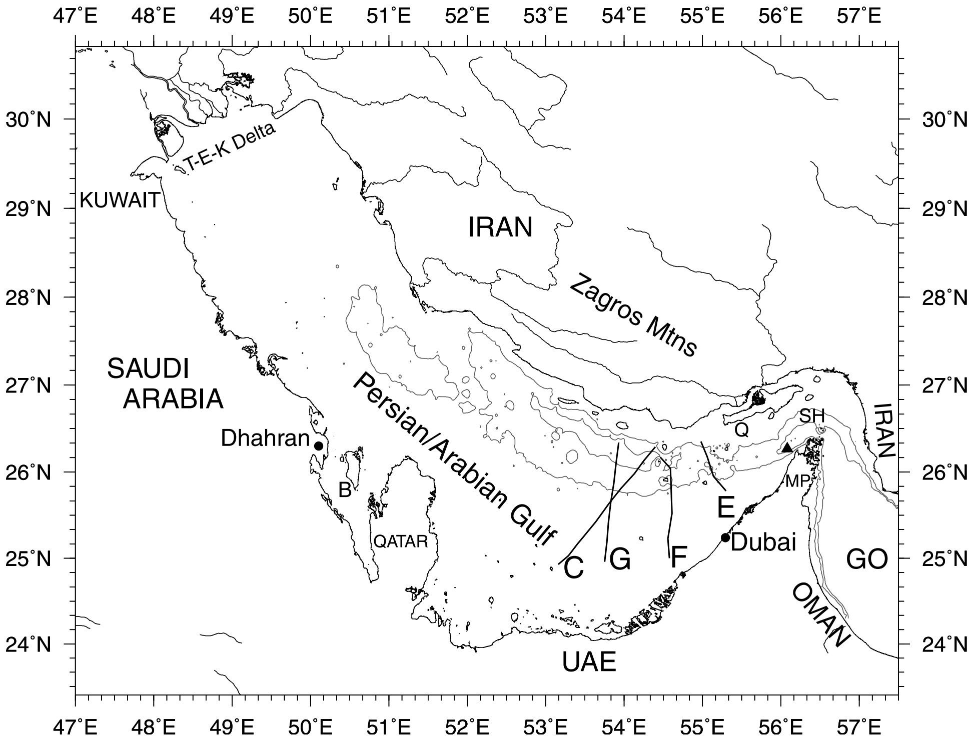

Figure 1. The Persian/Arabian Gulf (Gulf) is separated from the Gulf of Oman (GO) by the Strait of Hormuz (SH). Qeshm Island (Q) borders the north side of the Strait. Bahrain (B) lies between Qatar and the Saudi Arabian coast. The Tigris-Euphrates-Karun River delta forms the northern end of the gulf. Transects labeled E, F, and G were collected during the1992 *Mt. Mitchell* cruise in February and May. Transect C was collected during the February 1977 *Atlantis* II cruise. The triangle in the strait just west of the Musandam Peninsula (MP) marks the location of the mooring deployed by *Johns and Olson* [1998]. The 60 and 80 m bathymetric contours are shown.

although the opposite was true in late February 1992 [*Reynolds*, 1993]. Indian Ocean Surface Water (IOSW) normally flows into the Gulf from the open ocean along the northern side of the Strait and continues northward along the Iranian coast [*Emery*, 1956; *Brewer et al.*, 1979; *Hunter*, 1983; *Reynolds*, 1993], but this current appeared banked against the Oman (south) side of the channel in April, 1977 (Sonu in the work by *Chao et al.* [1992]). Based on estimates of evaporation-precipitation and the bulk temperature (T)-salinity (S) properties of the inflow and dense outflow through the Strait, *Hartmann et al.* [1971] and *Ahmad and Sultan* [1991] computed an annual mean exchange of 0.1–0.2 Sv. This estimate agrees fairly well with 0.28 Sv that *Johns and Olson* [1998] computed based on an ADCP mooring in the Strait.

[5] Direct observations of the circulation within the Gulf are scarce. Analysis of ship drift records indicates northwest flow with speeds greater than 10 cm/s along the Iranian coast from the Strait to the change in trend of the coast near 51.5°E and southwest flow in the regions of the southern Gulf away from Iran [*Hunter*, 1983; *Chao et al.*, 1992]. During the *Mt. Mitchell* cruises in early 1992, currents measured by vector-averaging current meters at 10 and 30 m depth on one mooring (mooring M2 [*Reynolds*, 1993; *Abdelrahman and Ahmad*, 1995]) agree with the historic ship drift data, but shallow meters on other moorings in the low salinity IOSW current indicated slow (<3 cm/s), southwest flow (mooring M3). Movement of drifter buoys deployed in the northwest current off southern Iran during the same period moved only 1–3 cm/s when the vectors are summed over the same period. Thus ship drifts, current meters on one mooring, and changes in the salinity field suggest northward flow at 4–5 cm/s for the Iranian half of the Gulf south of 27.5°N, whereas current meters and Lagrangian drifters at nearby locations suggest weaker flow to the west or south. The reason for this discrepancy is unclear, but the drift of the buoys, which were drouged at 0–1.3 m, may have been influenced by strong winds blowing from the northwest as well as the northwest-flowing IOSW coastal current.

[6] Modeling suggests that strong, northwest winds in the winter and spring produce southeast-flowing surface currents along both coasts in the northern gulf, confine cyclonic circulation to the southern Gulf, and shift the surface current through the Strait to the south side of the channel [*Chao et al.*, 1992; *Lardner et al.*, 1993]. Based on seasonal cycles in solar heating, wind, and evaporation and on cross-sections across the Strait, *Chao et al.* [1992] suggested that inflow through the Strait peaked at 0.17 Sv in March and decreased to 0.03 Sv in August–September. *Reynolds* [1993] found that isopycnals below 50 m water depth in a hydrographic section across the Strait in February–March, 1992, dip down to the south but dipped northward in a section in May–June and agreed that the exchange was seasonally dependent. *Johns and Olson* [1998], however, observed relatively constant outflow cur-

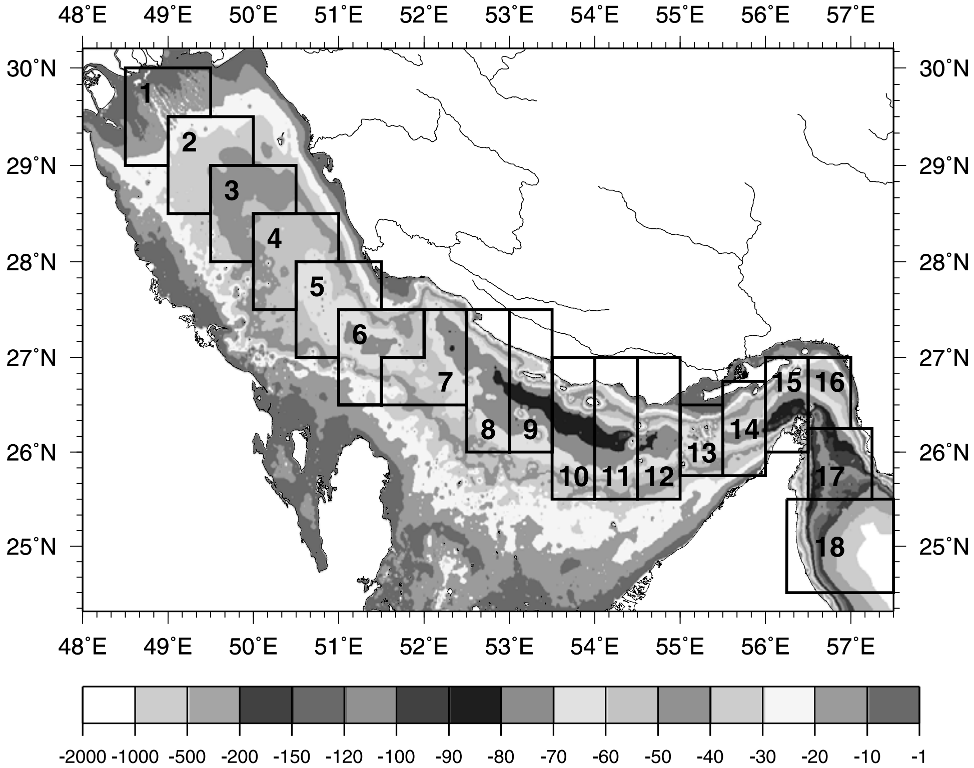

Figure 2. Bathymetry of Gulf compiled from navigation charts shows elongate Gulf basin narrowing in the western approach to the Strait of Hormuz. From shallow banks bordering the Arabian side of Gulf, the seafloor dips northward to an 80–105 m NW–SE trough near Iran. A sill at ~86 m occurs near 55°45′E between the main trough and the 100–200 m deep channel through the Strait of Hormuz. Another sill occurs southeast of the Strait at ~110 m near 26°N. Boxes indicate regions in which station data were obtained to form an along-axis vertical section. Stations on the shallow banks off Arabia were excluded from the axial section.

rent speed from an upward looking ADCP moored in the western channel (location shown in Figure 1). They also observed a dramatic shift in the T-S properties of water flowing through the Strait at most depths in July and a more gradual reversal in properties between November and January. It appears that the picture of circulation in the Gulf and exchange with the Indian Ocean from the modeling and the isolated surveys is not complete.

[7] A seasonally complete data set would help resolve these inconsistencies. Comprehensive surveys of currents and water properties throughout the year have not been obtained in the Gulf. In this paper we examine a hydrographic database and resolve seasonal trends by averaging the data from many cruises over many years. We identify the source areas and season when dense water is formed in the Gulf and use T-S characteristics to trace movements and transformations of water masses within the Gulf and the exchange with the Indian Ocean. With this approach we are able to base our interpretations on both the fine-scale spatial and temporal resolution of existing synoptic surveys and on bimonthly averages for about two thirds of the year.

2. Data

[8] Temperature and salinity values as a function of water depth were obtained from a database maintained by the U.S. Naval Oceanographic Office (MOODS, Master Oceanographic Observations Data Set). *Alessi et al.* [1999] describe general features of the data and the quality control procedures applied to edit out aberrant stations. Data from both bottle casts and CTDs at 1758 stations cover most areas of the Gulf west of 57°30′E. Figure 3 shows the distribution of stations as a function of month pairs. *Eid and El-Gindy* [1998] used a similar historical data set to study Gulf circulation but confined their analysis to geostrophic computation of surface currents.

[9] We chose MOODS as our primary data source rather than the National Oceanographic Data Center (NODC) because NODC does not include data from the 1992 *Mt. Mitchell* cruise, the most extensive synoptic survey of the Gulf [*Reynolds*, 1993]. The MOODS data include large temporal gaps and a bias toward winter and spring months (Tables 1 and 2). To fill these gaps we searched published and digital databases, including NODC. We obtained additional data collected prior to 1964 from *Dubach* [1964] and data collected on the 1977–1978 R/V *Lemuru* cruises from [*Simmonds and Lamboeuf*, 1981]. Neither MOODS nor NODC include data from the R/V *Umitaka-Maru* cruises during December and January of 1993 and 1994 [*Otsuki et al.*, 1998]. However, these data do not fill a significant data gap and would be unlikely to change our results.

[10] The data include spatial bias as well (Figure 3). The axial channel through the Strait and its approaches has been poorly sampled. This is somewhat surprising given the

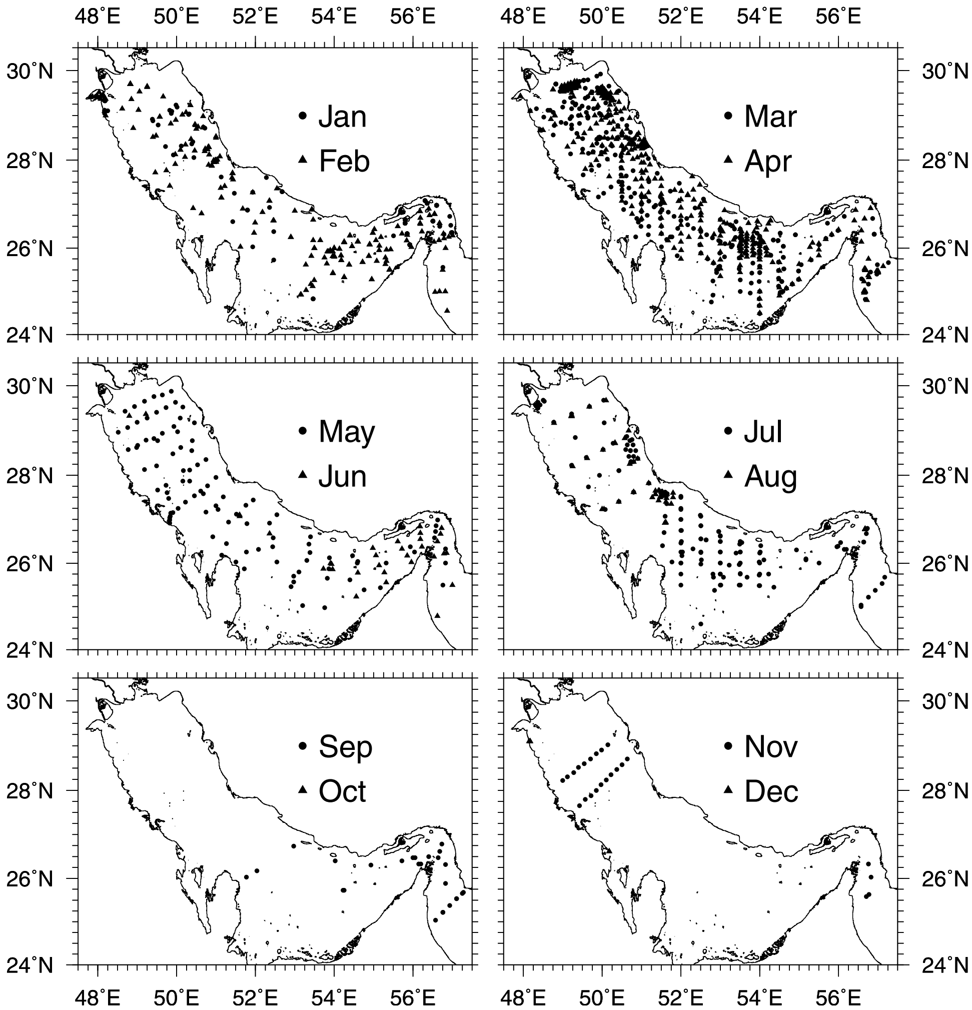

Figure 3. Locations of casts by seasons indicates few stations in summer months and almost no stations during the fall-early winter period. For most months there are too few stations in the Strait of Hormuz to adequately document the variability in water masses. With the exception of the February–June period, there are no stations on the shallow banks off the UAE southern coast.

obvious importance of this region to understanding the exchange with the Gulf of Oman but probably results from the difficulty of surveying in a major shipping channel and political difficulties. In addition, no stations are located within shallow bays and basins along the Arabian coast, and only a few stations, collected in March 1977 and the 1990s, are located within the topographic bank and channel complex linking the coastline basins to the deeper axis of the Gulf. As a result, very high salinity and density water found closest to the coast [i.e., *Sugden*, 1963; *Chandy et al.*, 1990] is not represented in our analysis. We suspect that these waters are trapped by topography and only reach the Gulf episodically.

Table 1. Distribution of Casts by Decade

Decade	Number of Casts
1940s	16
1950s	96
1960s	644
1970s	61
1980s	0
1990s	941

[11] Our data set includes a small number of synoptic surveys. The earliest is a large survey during July–August of 1968 (273 casts). The best documented survey is the R/V *Atlantis II* survey of the Gulf in February of 1977 [*Brewer et al.*, 1978; *Brewer and Dyrssen*, 1984]. The *Mt. Mitchell* surveys in winter (late February–early March) and spring (late May–early June) of 1992 included comprehensive CTD sampling that covered many regions of the Gulf that

Table 2. Distribution of Casts by Month

Month	Number of Casts
January	148
February	223
March	287
April	542
May	114
June	64
July	160
August	161
September	26
October	0
November	27
December	6

were poorly sampled by other expeditions, including the Strait of Hormuz and the shallow southern banks [*Reynolds*, 1993; *Sultan and Elghribi*, 1996]. Follow-up CTD surveys were done in April of 1994 and March–April of 1996, but the area investigated by these cruises is not as wide as that covered by the *Mt. Mitchell*.

3. Results

3.1. Formation of Dense Gulf Bottom Water

[12] Formation of dense water in the Gulf is commonly associated with processes that increase salinity [*Emery*, 1956; *Chao et al.*, 1992]. This viewpoint follows from the understanding of the Gulf as a shallow reverse-estuary system in an arid, subtropical setting. It is reasonable to presume a direct link between the production of high salinity water in particular regions and seasons with the formation of dense water that spills along the bottom through the Strait into the Indian Ocean. While a small portion of the flux out through the Strait is indeed hypersaline (>41 psu) water, we shall argue that most of the dense outflow is water that has been diluted by mixing in several steps before exiting the Gulf.

[13] The densest water in the main Gulf basin forms in winter. Figure 4 shows T-S plots for all data inside the Strait by month pairs. The highest density water, with $\sigma_\theta > 29.5$, is observed only between January and April. Most of this water has salinity less than 41 psu. We found 1258 samples with $\sigma_\theta > 29.5$ and obtained an average temperature (mean ± standard deviation) of 16.9 ± 1.5°C, salinity of 40.6 ± 0.5 psu, and a density of 29.8 ± 0.3 σ_θ units. The densest water is formed from surface water by cooling of warm mixed layer water by up to 13°C and evaporative salinity increases of up to 3 psu (Figure 4). This densest water mixes with relatively fresh, warmer inflow from the Gulf of Oman to form Gulf Deep Water, the most common water mass in the Gulf with modal peaks at about 19°C, 39.75 psu, and 28.25 σ_θ units for most months of the year (Figure 5). Formation of dense water may begin as early as November when air temperatures decrease and wind speed increases, but too little data are available for this period to describe the breakdown of the seasonal thermocline and the increase in vertical and lateral mixing that accelerates heat and water fluxes at the sea surface in early winter.

[14] Paradoxically, the densest water forms during the months of peak rainfall [*Kappus et al.*, 1978]. This results, in part, from the small magnitude of the flux from precipitation and, in part, from the delay of peak river discharge until the period from March to May when the snow pack melts in the Zagros Mountains of Iran and the Taurus Range of eastern Turkey [e.g., *Beaumont*, 1973]. In addition, river discharge from both the Shatt-Al-Arab and the Iranian rivers are diverted into currents flowing southward along both the Arabian and Iranian coasts [*Brewer et al.*, 1978; *Lardner et al.*, 1993; *Reynolds*, 1993] that reduce their immediate impact on water mass formation in the central portions of the Gulf.

[15] Most of the dense water forms above the shallow seafloor at the northwest end of the Gulf with lesser amounts forming along the lagoons and shallow banks off the southwest and southern coastlines. Figure 6 (top) shows the location of all the casts which had at least one sample

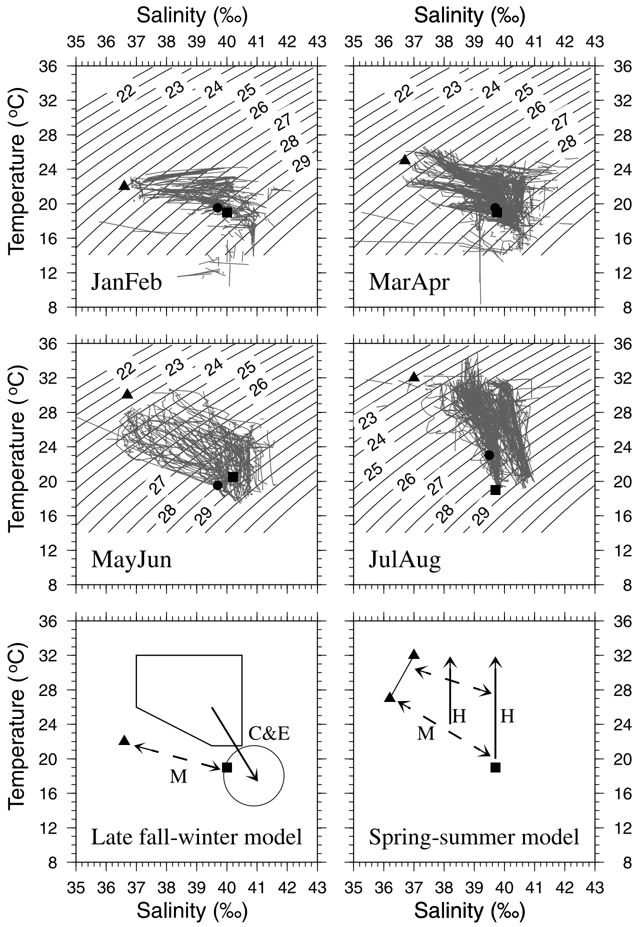

Figure 4. Data from all casts in Gulf indicate seasonal changes in T-S relationships. Only data within the Gulf (longitude <56.4°E) are shown. Fall months are omitted because there are too few casts to describe range of T-S characteristics. Mixing with IOSW (triangle) occurs throughout the year. From March to August, solar heating raises temperatures. Surface salinity increases in July–August. Highest densities occur in January–April. Bottom two panels depict schematically the principal processes affecting water masses in general. In late fall to early spring, surface cooling (C) and evaporation (E) of surface and intermediate depth water (square) produces (solid arrow) the densest water found along the northern and southern coasts (circle). In late spring and early fall, surface heating (H) warms surface and intermediate water (solid arrow). In all seasons, mixing (M) occurs between Gulf Deep water and IOSW (dashed arrows). Symbols indicate average T-S characteristics: circle is bottom outflow water in Strait (box 15 in Figure 2), square is modal peak of all water in Gulf (from Figure 5), and triangle is IOSW (depth <10 m, box 18 in Figure 2).

with density greater than 29.5. A few samples are found along the Arabian coasts, but the majority are found in the northern third of the Gulf off the Tigris-Euphrates-Karun delta. Only one station collected a high-density sample in the western approach to the Strait of Hormuz. Axial cross-sections obtained by averaging data in bins (Figures 7a, 7b,

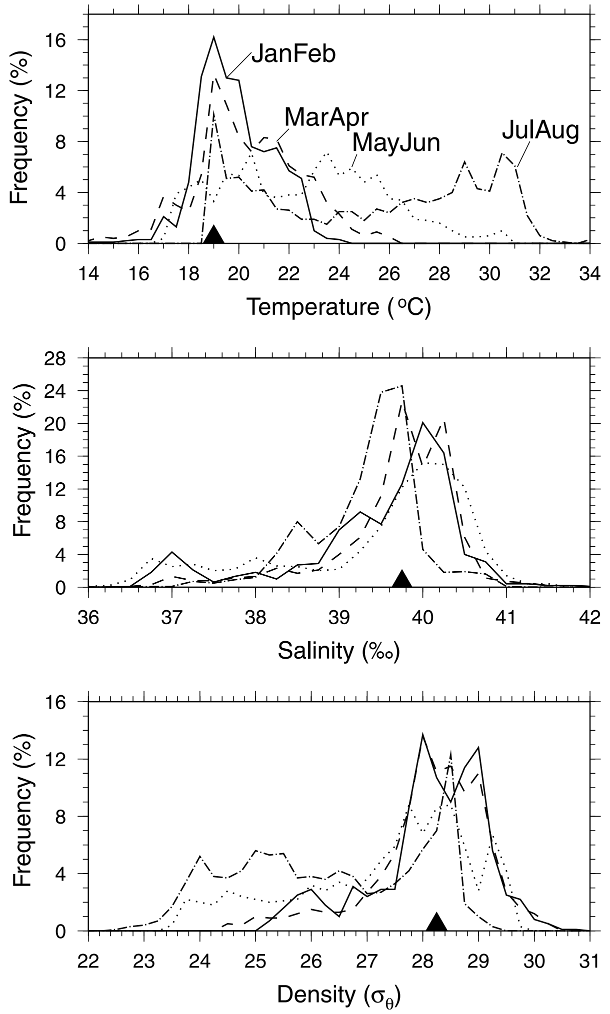

Figure 5. Numerical census of observations indicates modal water masses in the Gulf by bimonthly period. Triangles indicate locations of the principal mode for the entire year. The temperature census shows a sharp peak in most months at 19°C, the temperature of Gulf Deep water forming the bottom water layer in the interior of the Gulf. Higher temperature modes in May–June and July–August indicate surface water lying above the seasonal thermocline. The salinity census shows one principal mode that falls within the 39.5–40.4 psu range with no systematic changes with season. The density census also changes little with season, having a single, broad mode at 27.8–29.4.

7c, and 7d) clearly show that the densest isopycnals in the Gulf outcrop at the surface in the northern end of the Gulf from at least January to April, when northwest winds bring interior continental air to cool the surface of the Gulf. Air temperature over the Gulf is coldest in the north and increases southward by heat exchange with the sea surface [*Brower et al.*, 1992]. As a result, heat loss from the sea surface is greatest in the north, and the densest water forms there. In winter, wind speeds are higher and humidity is lower [*Perrone*, 1979; *Brower et al.*, 1992], contributing to higher evaporation rates than in summer. Although the coast is open and circulation is largely unrestricted, evaporation rates are high enough to increase salinity to 40–41 psu (Figures 4, 7a, 7b, 7c, and 7d).

[16] Dense water also forms along the Bahrain-Qatar shelf and behind the shallow banks off the United Arab Emirates (UAE) along the south coast. Fewer numbers of high-density samples occur in the south than in the north (Figure 6, top). Part of the reason may be sampling bias; there are fewer samples in the south (Figure 3). Warmer temperature is also part of the reason. High-density water is found along the south coast only between January and March, and winter weather is warmer here than at the north end of the Gulf. Tables of *Brower et al.* [1992, pp. 249, 251] indicate that daily mean air temperatures in January range from 15–23°C and average 19°C in Dubai, UAE, whereas the range in air temperature is 6–17°C and the average 13°C in Kuwait. As a result, those water samples with high-density (σ_θ > 29.5) south of 27°N latitude (chosen arbitrarily) have higher average temperatures (18.7 ± 1.6°C) than do high-density samples in the north (16.8 ± 1.4°C). The contribution of winter cooling to dense water formation is not as great along the south coast as in the northern Gulf.

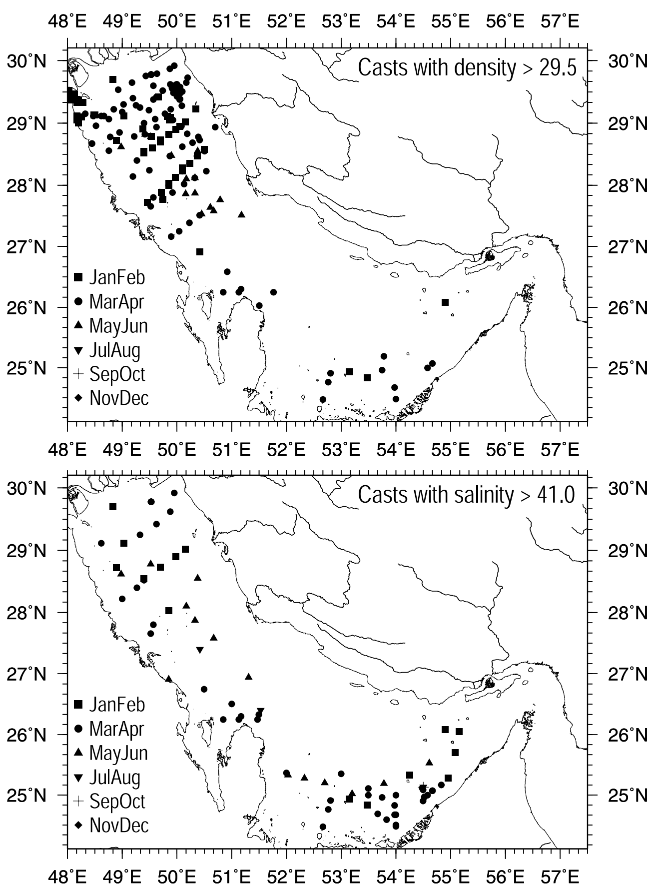

Figure 6. Most stations with at least one sample having density >29.5 occur in the northern end of the Gulf, and most were collected in the January–April period. High salinity stations (S > 41 psu) are equally distributed between the northern Gulf and the banks along the south coast. Almost no stations in the axial trough or the western approach to the Strait have high density or salinity.

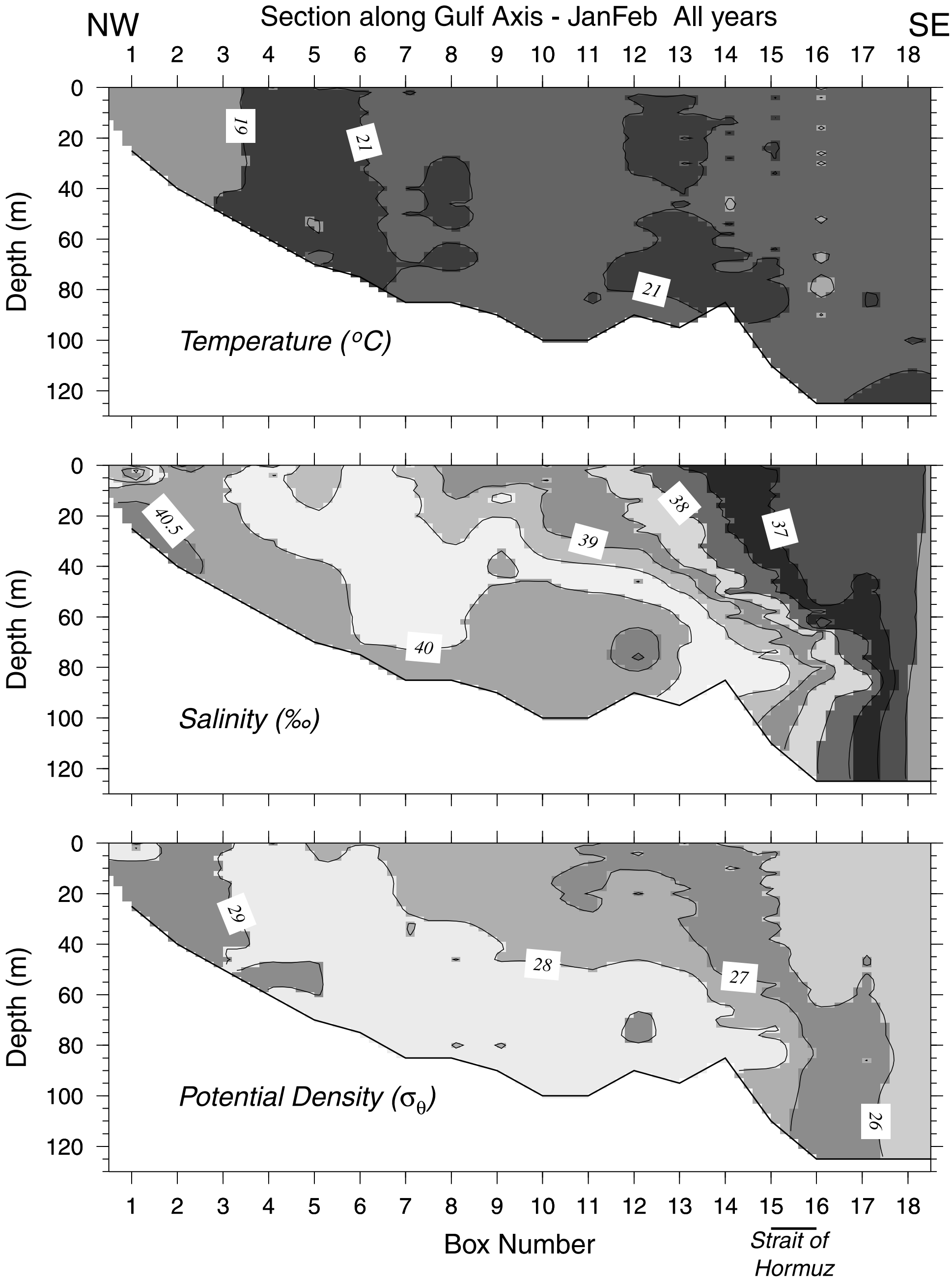

Figure 7a. Section for January–February along deep axial basin from Gulf of Oman (right, box 18) to Tigris-Euphrates-Karun River delta off Iraq (left, box 1) shows most of the high-density water at the head of the Gulf. Dipping front between modified IOSW (above) and Gulf Deep Water (below) is clear in salinity and density sections. Data from all years in data set were averaged in 2 m depth bins. Figure 2 shows location of boxes. Local density inversions are likely an artifact of binning irregularly sampled data.

[17] High-density water in the south is more saline than high-density water in the north. On average, samples with high density ($\sigma_\theta > 29.5$) south of 27°N latitude have higher salinity (41.6 ± 0.9 psu) than do high-density samples in the north (40.5 ± 0.3 psu). High-salinity values, in general, are more common along the south coast. Figure 6 (bottom) shows that samples with salinity values in excess of 41.0 psu are as common in the south as in the northern half of the Gulf, despite there being fewer samples overall in the south. We found little spatial variability in the salinity and density of high salinity samples. Samples south of 27°N with salinity greater than 41 psu have an average temperature of 22.6 ± 2.6°C, salinity of 41.6 ± 0.5 psu, and density of 29.0 ± 1.0 σ_θ units, whereas high-salinity water samples for the whole basin have a temperature of 19.1 ± 1.7°C, salinity of 41.5 ± 0.6 psu, and a density of 30.0 ± 0.6 σ_θ units. Extremely high salinity values (>50 psu) have been observed in shallow bays west of Qatar [*Sugden*, 1963;

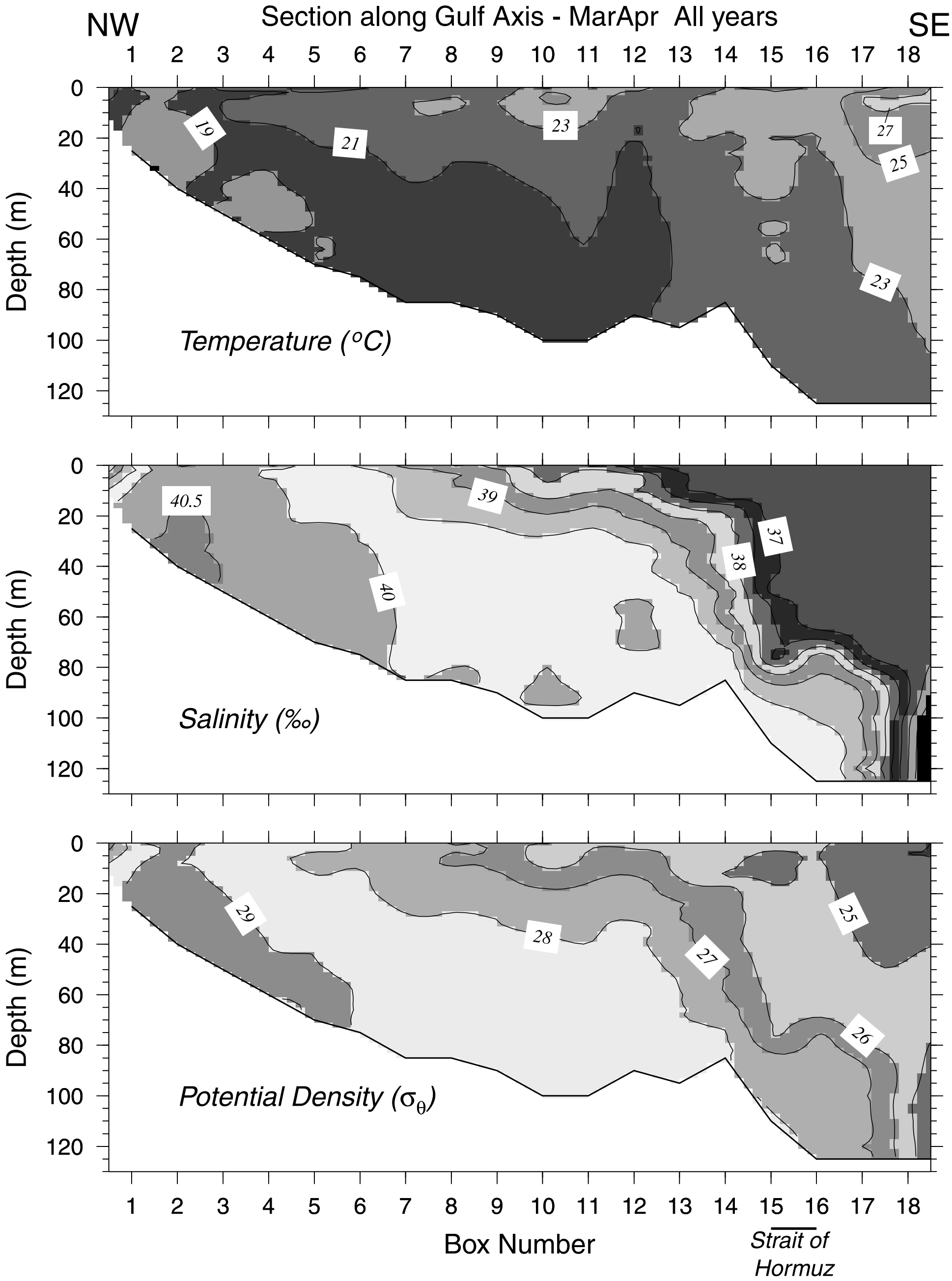

Figure 7b. Axial section for March–April shows cold, saline water still outcrops at the sea surface of the Tigris-Euphrates-Karun delta. Warm, low salinity IOSW at the sea surface has moved farther up the Gulf.

Chandy et al., 1990] and probably are common in lagoons elsewhere along the south coast. Such high-salinity values do not appear in our data indicating considerable dilution before reaching the shallow southern banks and the main Gulf basin where our samples were taken (Figure 3). High salinity water formed along the coast appears to be trapped there by the same topographic and oceanographic restrictions that allow salinity to increase to high levels.

3.2. Structure and Variability of Density Gradients in the Gulf

[18] A strong density gradient separates cool, saline water in most of the southern Gulf from warmer, fresher surface water that enters from the Gulf of Oman. This gradient usually takes the form of a flat or gently sloping pycnocline. During winter, the dip of isopycnals in the Strait and along the southern flank of the main basin is steeper, isopycnals crop out at the sea surface, and the gradient takes the form of a front separating two horizontally distinct water masses. The nature and dynamics of the gradient reflect and, to some extent, control exchange processes in and out of the Gulf. We describe the gradient here and later use changes in location and the characteristics of water masses across the gradient to interpret the nature of the exchange processes and their seasonal pattern.

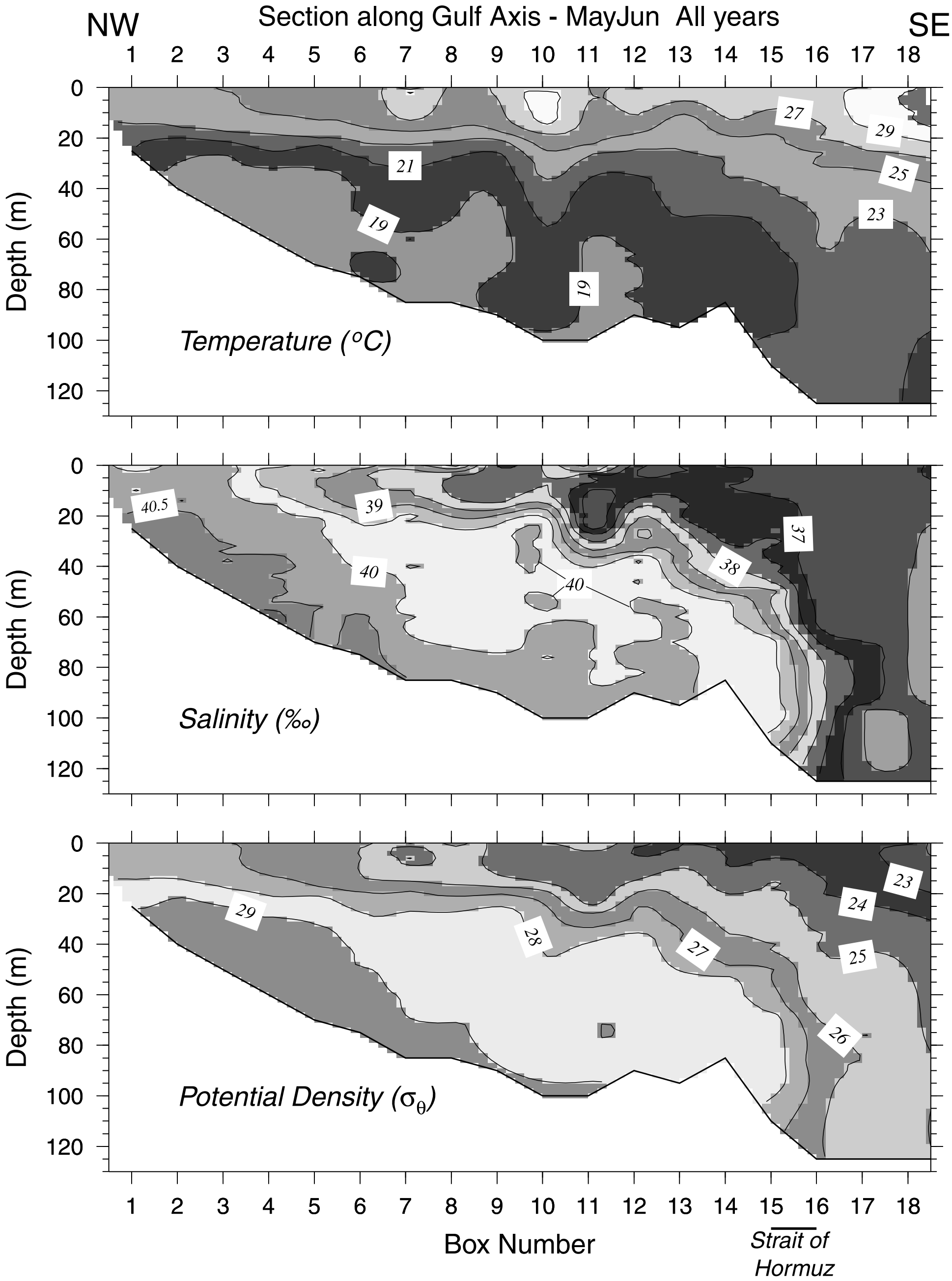

Figure 7c. Axial section for May–June shows that the development of the seasonal thermocline has cut cold, saline Gulf Deep Water off from sea surface. Salinity data indicate that modified IOSW at the sea surface reaches closest to the Tigris-Euphrates-Karun River delta during these months.

[19] In the Strait of Hormuz, the density surfaces appear to be near-vertical and oriented north-south across the Strait. Long density sections down the Gulf axis, collected during February by *Brewer et al.* [1978, Figure 20] and *Reynolds* [1993, Figures 11 and 12], show that isopycnals in the Strait dip steeply downward toward the Gulf of Oman and intercept the seafloor. Stations in these sections are too far apart to image the nature of the density gradient in the Strait itself, but isopycnals there appear to be near-vertical. Due to tanker traffic and political obstacles, publicly available data obtained in the Strait bend north of the Musandam Peninsula is too sparse to reliably describe the details of the water mass and density structure. *Matsuyama et al.* [1994, 1998] show the most detailed section along the Strait to be published. In agreement with *Brewer et al.* [1978, Figure 20], the σ_t = 25.5 isopycnal clearly outcrops at the sea surface, and water with $\sigma_t > 26.0$ appears at the seafloor inside the Strait (to the west) but not outside the Strait. This section, however, meanders in and out of the main outflow pathway, and it is unclear how well it images axial structure along the deepest part of the channel. In the cross-axis direction, other data indicate little variability. Isopycnals in a north-south section collected at 56.2°E during February 1992, put the base of the IOSW at 40–50 m, show little

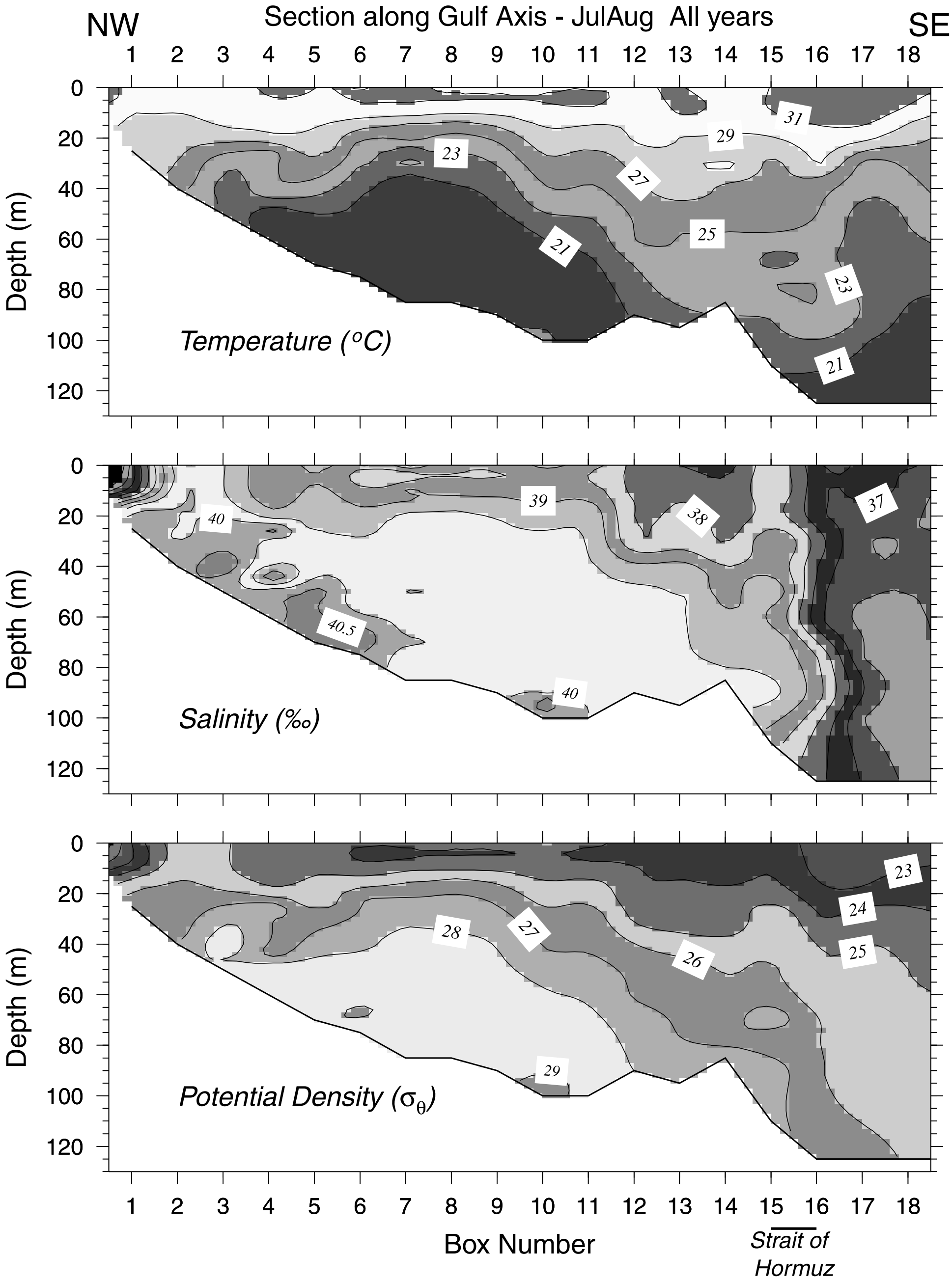

Figure 7d. Axial section in July–August shows that the southeast end of the front separating modified IOSW from Gulf Deep Water has moved up the basin to at least box 12. Warm, low-salinity surface water has been stirred downward in the region to the west of the Strait (boxes 12–15). Solar heating has warmed the complete water column lowering overall density compared to previous months.

change in depth across the Strait, and appear to intercept both sides of the channel [*Reynolds*, 1993, Figure 14]. Summarizing the limited data available, density surfaces in the Strait during winter appear to be planar features that are near-vertical at the bend in the Strait north of the tip of the Musandam Peninsula. Within the Gulf, density surfaces dip down toward the Gulf of Oman. Isopycnals in sections taken across the channel axis slope more in spring than in winter [*Chao et al.*, 1992; *Reynolds*, 1993]. Unfortunately, axial sections with stations as closely spaced as that of *Matsuyama et al.* [1994] have not been collected in the spring, and no data from the Strait with such high density and quality is yet available for summer months.

[20] Within the southern half of the Gulf west of the Strait, vertical sections down the axis of the main basin collected during February show isopycnals sloping from the seafloor upward to the west in a fan pattern [*Brewer et al.*, 1978, Figure 20; *Reynolds*, 1993, Figure 11]. As a result, isopycnals are spaced farther apart at the sea surface than at the seafloor. Isopycnals outcrop at the sea surface in the interior of the basin, so maps of surface density and salinity show horizontal gradients that mark the surface location of

the water mass boundary [*Brewer et al.*, 1978, Figures 4 and 6; *Brewer and Dyrssen*, 1984, Figure 3; *Reynolds*, 1993, Figure 9]. The gradient separates modified IOSW to the northeast along the Iranian coast from cooler, more saline Gulf water to the southwest. The southern end of the front can traced back to the UAE coast near 55°E where a section by *Brewer et al.* [1978] (their Transect B in their Figure 9) shows the cluster of isopycnals marking the front separating from the southern wall of the channel and rising to the sea surface. The northern end of the gradient is less well-defined by data but probably does not extend north of 27°N.

[21] The water mass boundary changes with the season. In the spring, modified IOSW moves both farther up the Iranian coastline and southwestward along the UAE coast. As a result, isohalines outcropping at the sea surface spread northward up the Iranian coast and farther south closer to the UAE coast [*Reynolds*, 1993, Figure 10]. In summer, the gradient does not reach the surface. A basin-wide seasonal thermocline forms with a warm, well-mixed surface layer comprised of a mixture of IOSW and saline Gulf water.

[22] Two lines of evidence suggest that the downward slope of isopycnals toward the east in the Strait represents a region of mixing between water masses and not a boundary between flows moving in different directions. Axial density sections collected in winter clearly show that isopycnals intersecting the sea surface dip down toward the east into the Strait and pinch out at the seafloor [*Brewer et al.*, 1978, Figure 20; *Matsuyama et al.*, 1994, Figure 5; *Reynolds*, 1993, Figures 11 and 12]. In north-south sections across the western approach to the Strait, the density gradient at the base of the modified IOSW is near the sea floor [*Brewer et al.*, 1978, Figure 9; *Reynolds*, 1993, Figures 14 and 19]. If near-bottom outflow were constrained to the region beneath the pycnocline, the velocity of the outflow would have to increase many fold eastward to conserve volume. Current meter records in the Strait show no evidence that the amplitude of the mean flow increases near the sea floor [*Matsuyama et al.*, 1994, 1998; *Johns and Zantopp*, 1999], so it is likely that little Gulf water flows unmodified beneath the IOSW. The along-strait density gradient is a manifestation of the water mass transformation that occurs as the deep Gulf water passes through the Strait and mixes with the overlying inflow water. This inference is supported by the observation of *Johns and Olson* [1998] that outflow in the western approach to the Strait remained relatively constant from the seafloor up to ~40 m depth year-round, whereas the temperature and salinity profile at the mooring changed significantly with seasons. If the gradient were a flow boundary, measured outflow should have changed when the density profile changed. We will argue later that exchange through the Strait is driven by evaporative water losses at the sea surface in the Gulf and the density imbalance between Gulf bottom water and water outside the Gulf.

[23] The depth and location of density gradients give some evidence of how dense water over the shallow banks bordering the southern Gulf coast interacts with deep Gulf water. Leakage of salty, dense water off the banks appears to occur only during special oceanographic conditions and may be limited spatially and temporally. Transects across the Gulf in 1977, 1992, 1994, and 1996 indicate that the interface between modified IOSW and dense Gulf water in the deep basin is often nearly flat during late winter (Figure 8a). To the south of the basin axis, the pycnocline intersects the seafloor near the 40–50 m isobaths and then bends upward, forming a front that outcrops at the sea surface. Under these conditions the dense water on the banks is isolated from the deep basin. Under different conditions, the front slopes upward from the basin axis to outcrop at the sea surface above the banks without intersecting the seafloor. With the density front in a shallow position, there is a direct connection for the salty water to reach the axial basin. This condition occurred along transect F (~54°15′E) of the *Mt. Mitchell* survey in February of 1992 (Figure 8b) and along Transect C, collected oblique to bathymetry (but perpendicular to the front) in February 1977 (location in Figure 1; sections shown in Figures 10 and 11 of *Brewer et al.* [1978] and Figures 5–6 of *Brewer and Dyrssen* [1984]). These sections show salty, dense water flowing along isopycnals into the basin. The T-S plots for these sections (not shown) indicate lateral mixing between the dense bottom water and modified IOSW above the pycnocline. In contrast, *Mt. Mitchell* transects E (not shown) and G (Figure 8a), taken across the UAE margin to the east and west of Transect F, show dense shelf water trapped on the bank. It seems likely then that seepages of dense water off the bank in winter (January–March) are localized to regions with along-shelf widths of less than 100 km. The data indicate that conditions for basinward flow and mixing are more frequent in spring. Isopycnals parallel to the seafloor are observed in all four north-south transects across the UAE margin in April, 1994 (sections not shown). At this time, though, temperatures on the bank are much higher (22.5–25°C), and density is somewhat lower (0.5–1 unit). By the time of the *Mt. Mitchell* transects in late May 1992, most of the water on the banks is warm, modified IOSW, and salty water formed in winter and early spring has slipped into a pool on the seafloor above the 30–50 m isobaths (for example, Transect F in Figure 8c). The density of water in this pool is too low to mix downslope with Gulf deep water lying above the seafloor in the axis. This salty water appears to evolve along at least two pathways. A small portion mixes laterally into the basin beneath the seasonal thermocline and may be the source of 5–10 m thick high salinity layers observed at 20–30 m depth in small regions (40–80 km across) during the summer. A larger portion appears to be transported eastward along isobaths, appearing as a salty tongue of water at depths <45 m along the south wall of the Strait in summer [*Johns and Zantopp*, 1999] and as shallow, narrow veins of high T-S water in the Gulf of Oman [*Bower et al.*, 2000].

3.3. Spreading of IOSW

[24] Relatively fresh surface water from the Oman-Iran shelf flows into the Gulf (Table 3) to replace water lost to evaporation. As mentioned earlier, input from river discharge and precipitation is much lower than estimates of evaporative flux [*Hartmann et al.*, 1971; *Reynolds*, 1993]. For a long time oceanographers inferred that the surface water of the Gulf, in general, is saltier in winter than in summer. *Schott* [1908] attributed the difference to changing river fluxes, whereas *Emery* [1956] attributed the difference to a change in evaporation rate. Assuming that the evaporation rate increases when the air temperature

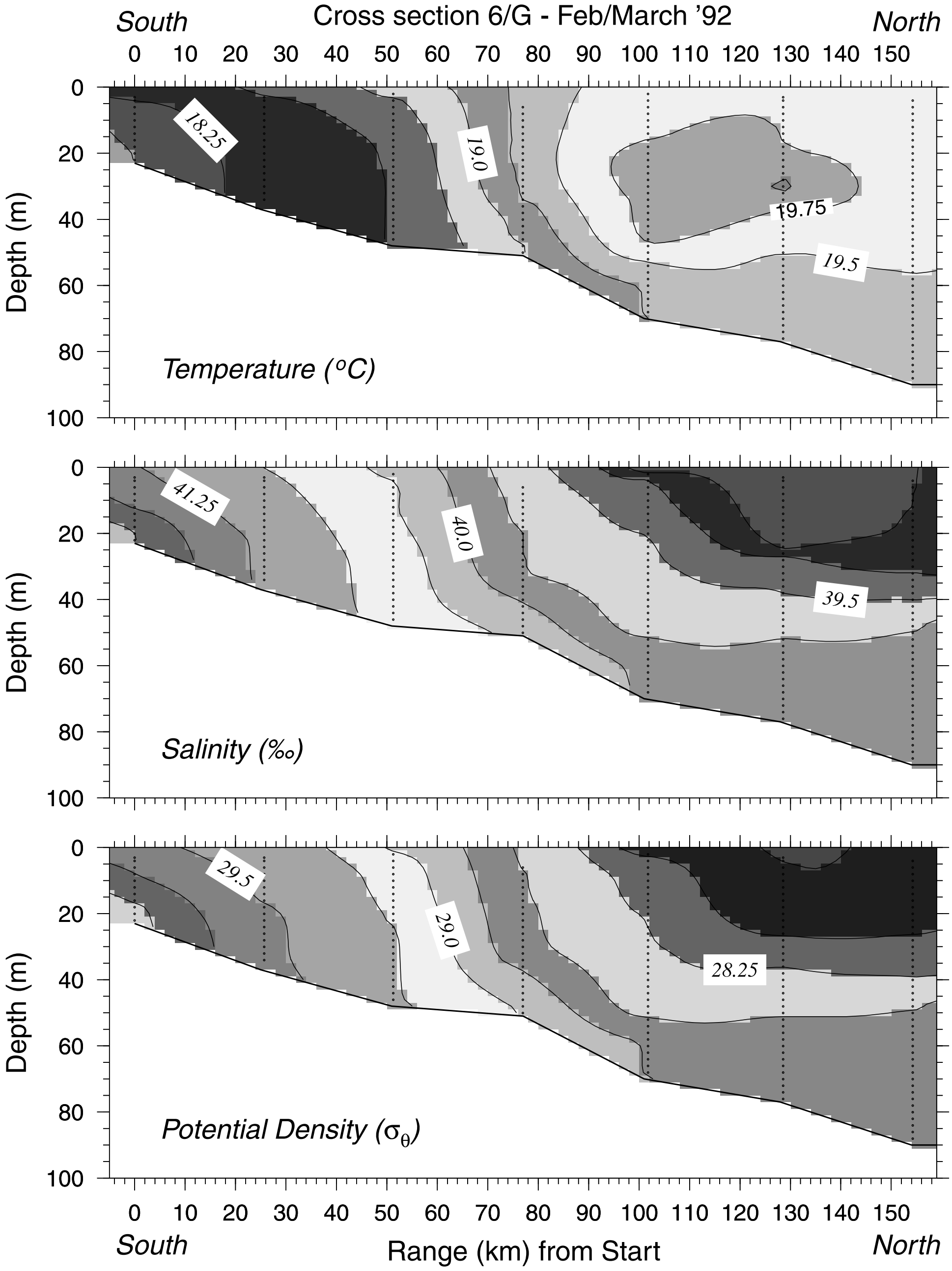

Figure 8a. Section G across the Gulf, obtained in February 1992, off the UAE (location in Figure 1) by the *Mt. Mitchell* [*Reynolds*, 1993], shows warm, low-salinity modified IOSW at the sea surface in the main axis of the basin. The front separating this water from denser Gulf deep water below bends upward above the sloping seafloor bounding the south side of the basin. This front blocks relatively dense, cold, high-salinity water formed on the banks off UAE from flowing into the basin and mixing with Gulf Deep Water.

dropped below that of the sea surface, Emery argued the evaporation rate for the Gulf would be higher in winter and would produce higher sea surface salinity. *Chao et al.* [1992] noted that the prevailing northwest wind is stronger in winter than in summer and suggested that the summer influx of IOSW is due to relaxation of surface wind stress. Our data suggest revisions in these interpretations are warranted. We use bimonthly maps and sections to show that influx of IOSW during some years may peak in late spring rather than summer and argue that the transport is driven by an evaporative lowering of sea surface height in the Gulf.

[25] Our data suggest that the presence of unmodified IOSW in the Gulf increases from January to June and then decreases in July and August. Maps of surface salinity (Figure 9) show that salinity less than 37 psu spreads westward into the Gulf from January–February into March–April. In May–June most contours spread 20–40 km further northwestward into the Gulf. Low salt water remains within the Iranian half of the Gulf despite a wide-

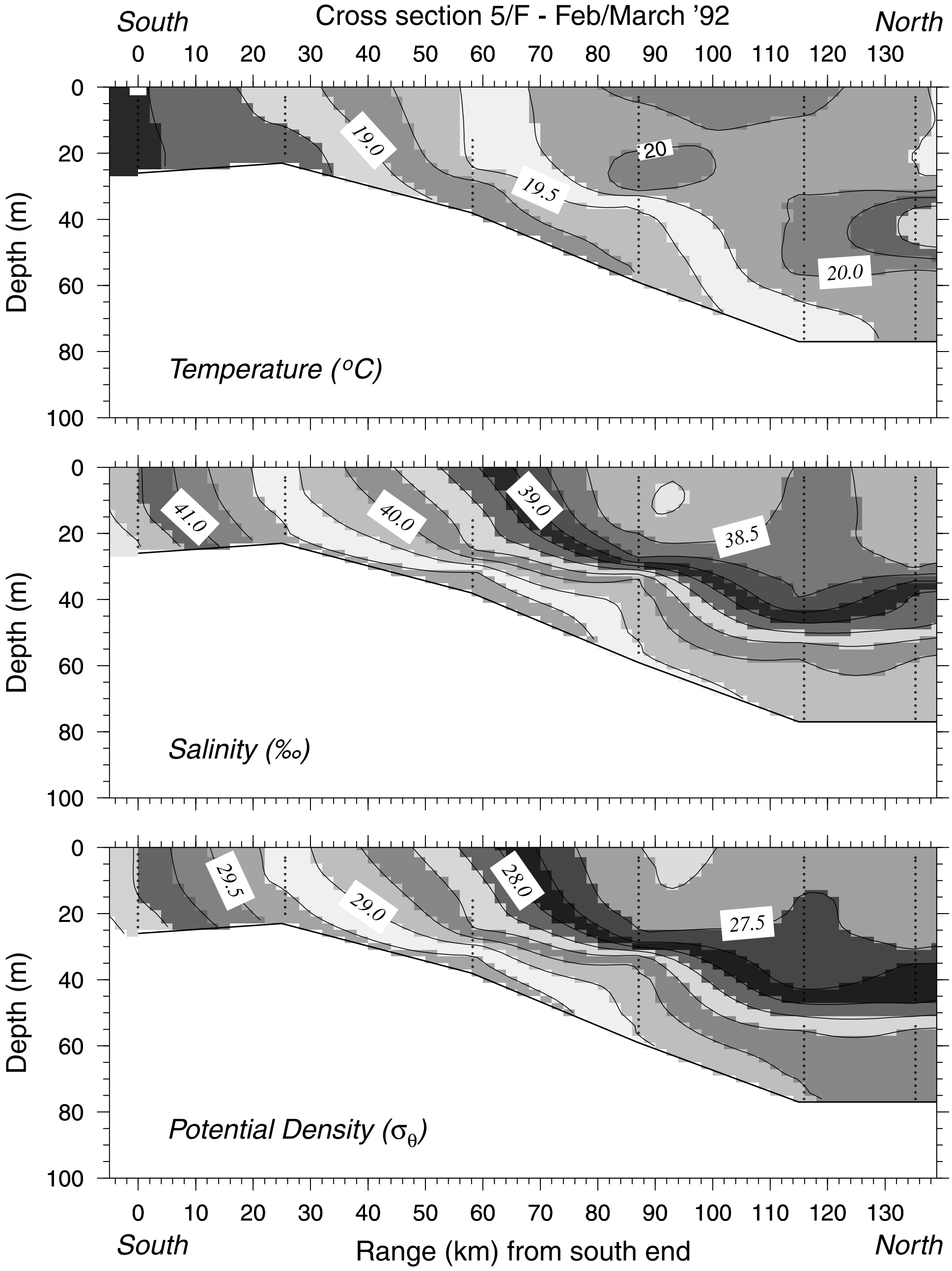

Figure 8b. North-south section F across the Gulf obtained in February 1992 40–50 km east of section G (location in Figure 1) by the *Mt. Mitchell* shows the front dipping at a more shallow angle than in section G. This section closely resembles that collected by the *Atlantis II* in February 1977 [*Brewer et al.*, 1978; *Chao et al.*, 1992]. Modified IOSW mixes onto the shallow bank at the sea surface and dense, saline water on the bank appears to flow down-slope to mix with Gulf Deep Water below the front. More exchange between shallow bank and deeper basin water may be occurring in this section than in section G to the west. Section D (not shown) to the east, however, shows that cold, saline, dense water is confined to the crest of the bank, similar to section G, so exchange at this time is limited to a 80–100 km section of the UAE coast that is crossed by section F.

spread, well-developed thermocline at this time. By July–August, however, water with salinity less than 38.0 psu retracts almost 100 km southeastward along the coast and covers an area smaller than it does in January–February. Surface salinity values greater than 40 psu have disappeared, and the distribution of values is more uniform than for other months.

[26] Along-axis sections (Figures 7a, 7b, 7c, and 7d) show a similar pattern. Water with salinity less than 38.0 spreads northwestward at the sea surface from box 12 during January–February to box 10 in March–April, about 110 km. Relatively low salinity water moves further northwestward to box 8 in May–June, another 190 km. In July–August, however, most of the water

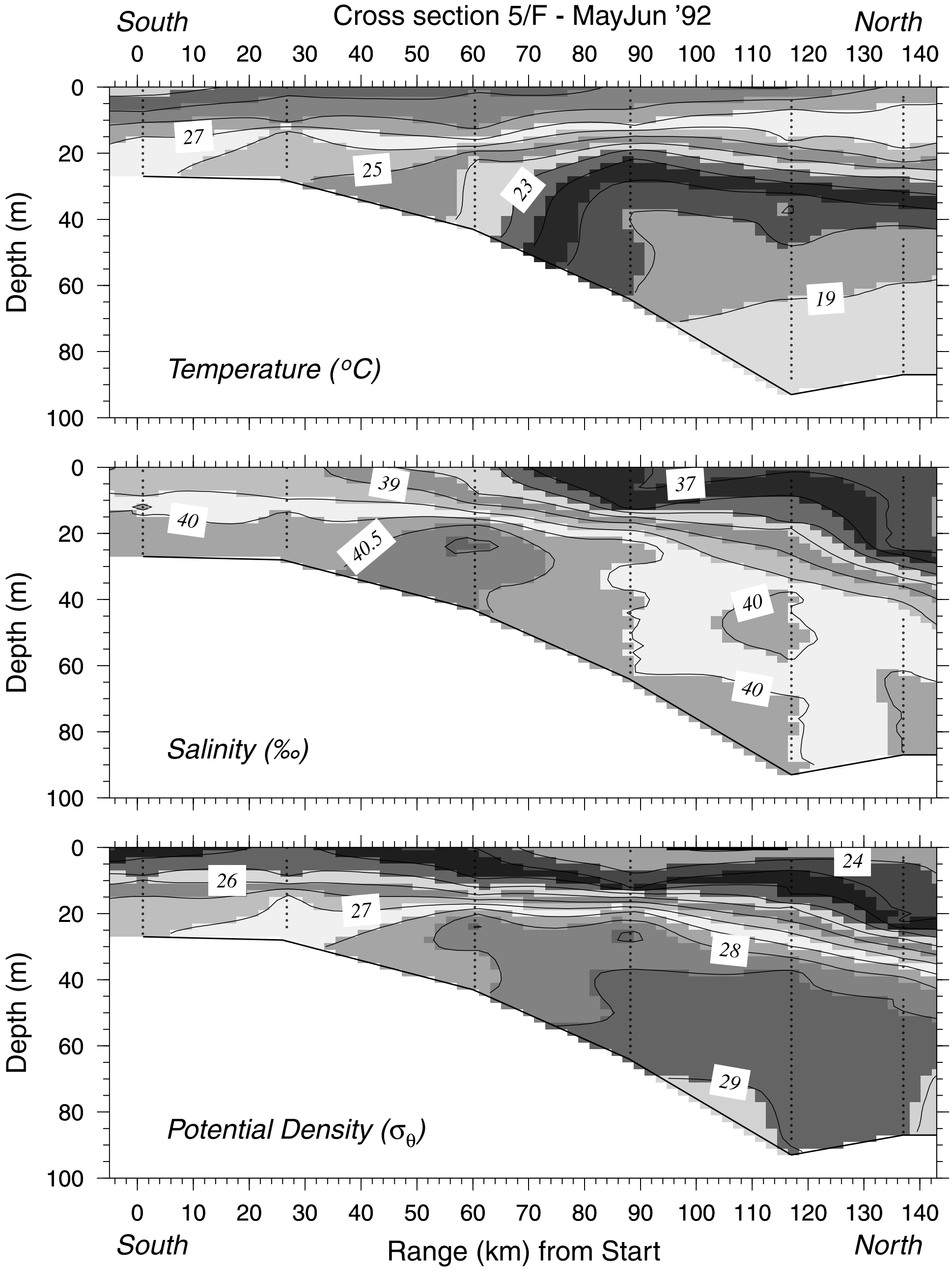

Figure 8c. North-south section F across the Gulf obtained by the *Mt. Mitchell* in late May 1992 at the same location as the section in Figure 8b shows the front dipping at a more shallow angle than in section G. Color scales differ from those used in 8a and 8b. High-salinity water formed on the bank in winter is now perched on the seafloor slope and mixes laterally with water in the dipping front. Dense Gulf Deep Water is isolated from exchange with the shelf at this time.

with salinity less than 38.0 has retracted south of box 12, a distribution similar to that in winter. The disappearance of IOSW in summer is apparent in T-S plots for the month pairs, as well. In Figure 4 temporal trends in surface salinities are difficult to resolve between January and June with most values falling in the 37–38 psu range. In July–August, however, minimum salinities clearly have shifted to values greater than 38 psu. These results appear to contradict the traditional interpretation that low-salinity IOSW is more widespread in the Gulf than in winter.

[27] Intermediate salinity values in our July–August data may not represent the only condition possible during the summer in the Gulf. The July–August data came from cruises in only 1968 and 1993. In both years the southwest-northeast salinity contrast across the Gulf in July–August is lower than during any other month pair, and salinity values less than 38 psu are found only as far west as 55°E. Salinities in 1968 appear more abnormally high than salinities in 1993. On the other hand, the surface salinity map from Emery's August 1948 survey shows a pattern more closely resembling our map of May–June than our map of July–August.

Table 3. Average Characteristics for IOSW (Sample Depth Less Than 10 m) by Month Pairs for the Strait of Hormuz (Boxes 15 and 16) and the Oman-Iran Shelf Outside the Strait (Box 17)[a]

	Temperature, °C		Salinity, psu		Density (σ_θ)		Number of Samples
	Mean	SD	Mean	SD	Mean	SD	
			Strait of Hormuz				
Jan.–Feb.	22.7	1.2	36.8	0.2	25.4	0.5	113
March–April	23.5	0.8	36.7	0.1	25.1	0.2	37
May–June	27.6	1.1	36.8	0.2	23.9	0.3	43
July–Aug.	32.0	0.7	37.2	0.3	22.7	0.4	85
			Oman-Iran Shelf Southeast of Strait				
Jan.–Feb.	21.9	0.7	36.6	0.1	25.5	0.2	27
March–April	26.6	1.2	36.8	0.1	24.2	0.3	72
May–June	29.6	1.0	36.9	0.2	23.3	0.3	31
July–Aug.	32.1	0.5	37.1	0.1	22.6	0.2	4

[a]See Figure 2 for location of boxes.

Apparently, the sea surface salinity distribution in July and August may vary significantly from year to year.

[28] Based on the MOODS data set, low salinity water (<37 psu) water reaches farther into the Gulf during the spring than in winter and summer. We can identify no simple explanation for the changes in surface water salinity other than a change in the rate of inflow of IOSW. In coastal Iran, rainfall peaks in mid-winter [*Kappus et al.*, 1978], but

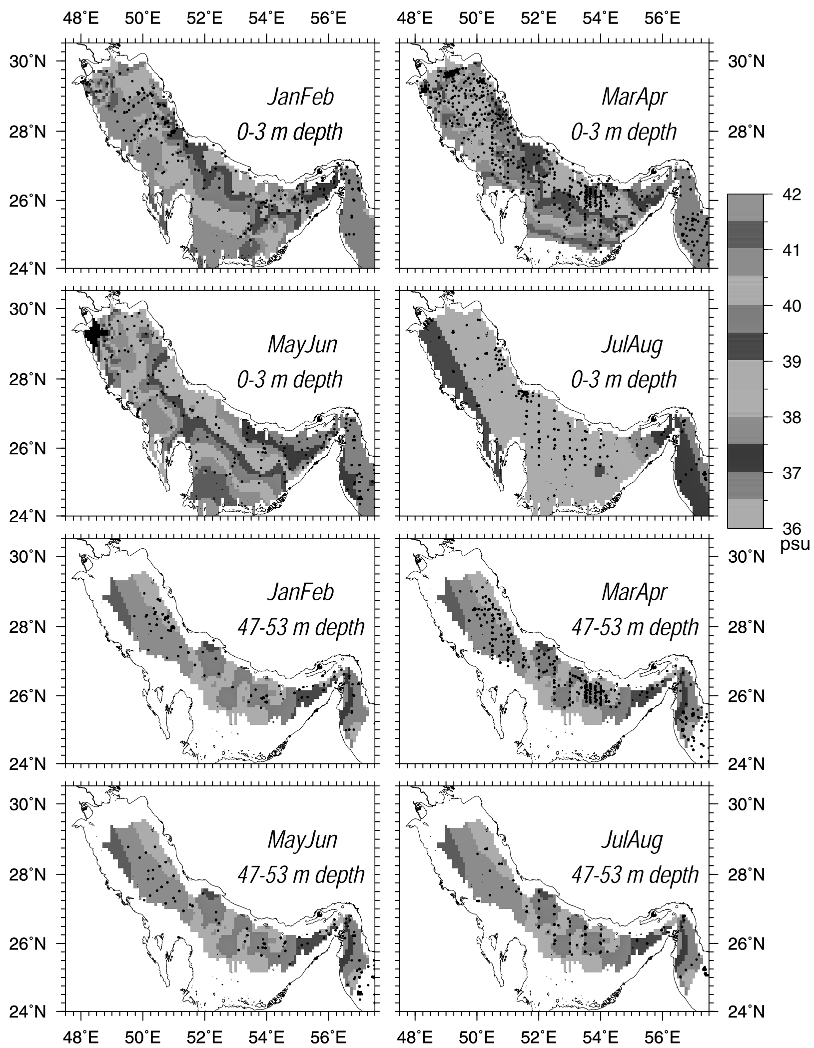

Figure 9. Surface salinity maps (top four panels) show that modified IOSW progressively moves farther up the Gulf from January to June. These data suggest that surface water with salinity less than 38 psu then retreats ~100 km eastward toward the Strait in July. Since the seasonal thermocline develops in May–June (Figures 5 and 7), this retreat is unrelated to summer warming. Salinity distribution of Gulf Deep Water at 50 m depth (bottom four panels) shows comparatively little seasonal change. See color version of this figure at back of this issue.

river runoff from the Zagros Mountain peaks during March–May when most of the snow melts [*Beaumont*, 1973]. The section of Iranian coastline along which low salinities are observed, however, does not have significant river discharge due to low annual precipitation in the mountain watersheds [*Beaumont*, 1973], so the salinity change is unlikely to arise from changes in fresh water runoff. In surface water samples (0–10 m depth) southeast of the Tigris-Euphrates-Karun coastline, we find no salinity values lower than that of IOSW on the Iran-Oman shelf at any time of the year, indicating that the discharge from other Iranian rivers has little affect on water properties beyond the coastal zone. Monthly average and peak wind speed remains nearly constant from December through June with strengths actually peaking in June at some stations [*Perrone*, 1979; *Brower et al.*, 1992; *Chao et al.*, 1992]. Wind speed drops during the July–October period. Since the prevailing surface winds blow from the northwest against inflow, the increase in low salinity surface water in the Gulf during the spring can not result from a decrease in surface wind stress, and the apparent recession of IOSW from the Gulf in July does not correlate with an increase in wind stress. We conclude that broader extension of low salinity values at the sea surface up the Iranian side of the Gulf indicates greater transport of IOSW through the Strait in late winter and spring.

[29] The apparent uniformity in surface salinity in July–August (Figure 9) could result from mixing water from high and low salinity sources, but a mechanism for increasing the lateral mixing rate between June and July is unclear. The uniformity could also result from lower offshore flow of high-salinity water along the southern coast as peak winds decrease and the number of severe northwest Shamal wind events decreases in July and from slower spreading of IOSW through the Strait. The available data are insufficient to reliably interpret summer surface conditions, and additional studies would be necessary to test these hypotheses.

[30] At least two mechanisms based on conservation of volume could explain the observed changes in IOSW transport. *Sultan et al.* [1995] argue that the annual change in sea level recorded by tide gauges in the Gulf (26 cm higher in summer) can be accounted for by changes in atmospheric pressure and surface water density (thermal effect, principally). Thus changes in the volume of the Gulf from other sources must remain minor, and fluxes through the Strait on a seasonal basis must balance. The seasonal changes in surface transport could be compensated by a seasonal change in outflow. The limited data available for deep outflow, however, suggest otherwise. *Johns and Olson* [1998] found a "relatively steady deep outflow" that varied little on a seasonal timescale. Even if this interpretation is in error by 10–20% due to changes in flow outside the channel axis, the magnitude of the changes in flux allowed are insufficient to account for the seasonal changes in volume of IOSW within the Gulf that we observe. There is a timing problem as well. The flux of IOSW appears to peak in late spring, whereas outflow of dense water would be expected to peak in winter when Gulf Deep water outcrops at the sea surface in the northern third of the Gulf (Figure 5) and the densest water ($\sigma_\theta > 29.5$) is most common (Figure 6).

[31] Alternatively, the surface influx could be driven by loss of water volume through evaporation. In this hypothesis, seasonal changes in water loss through the sea surface are compensated by changes in the flux of IOSW through the Strait. This is difficult to confirm because seasonal changes in evaporation rates in the Gulf are uncertain. *Privett* [1959] computed evaporation rate from shipboard meteorological observations, and found a peak during October–January and a minimum in May. This pattern is consistent with Emery's suggestion that the rate depends on the relative temperature of air and seawater. *Ahmad and Sultan* [1991] obtained a different seasonal pattern. They computed surface heat fluxes from meteorological observations at Dhahran, Saudi Arabia, and coastal sea surface temperature at stations farther up the coast. Their expression for latent heat flux is just the product of a constant and the evaporation rate. Their estimate of evaporation peaks between June and August, essentially the opposite result of Privett. The timing of this peak is within two months of the late spring intrusion of IOSW. This timing difference could be due to bias introduced by either the use of data from a single station to represent the whole Gulf or the use of onshore meteorological data to represent conditions just above the sea surface. Peaks in wind speed and minimums in relative humidity occur in May and June for many coastal monitoring stations [*Perrone*, 1979; *Brower et al.*, 1992], consistent with higher evaporation in these months. Clearly, a new survey of evaporation rate and its seasonal and spatial variability needs to be done to resolve uncertainties in published interpretations.

3.4. Deep Water Outflow

[32] As discussed briefly above, *Johns and Olson* [1998] found relatively little seasonal change in the current speed and thickness of the outflow along the channel axis through the Strait of Hormuz. Speeds averaged 20–30 cm/s from about 45 m depth to the bottom boundary layer [*Johns and Zantopp*, 1999]. While flow elsewhere in the Strait may vary seasonally, it seems reasonable to assume that the total flux also remains relatively constant. This apparent lack of change contrasts with the large percentage changes known to occur in air and sea surface temperature, wind speed, and humidity that would be expected to produce a excess of dense, saline water in either winter or summer depending on the data used to compute evaporation rate (respectively, *Privett* [1959] and *Ahmad and Sultan* [1991]). Whereas surface conditions clearly change with season, our data indicates that neither the salinity nor the density of water below 30–35 m depth in most of the Gulf changes significantly from January to August. So, we suggest that outflow is driven by a density contrast between this deep water and water at comparable depths outside the Strait that varies less than 10–15%.

[33] Salinity of water below the surface layer in parts of the Gulf changes little during the year. Figures 7a, 7b, 7c, and 7d shows that the cross-section area of salinity values of 39.5–41 psu remains nearly the same from January to August in the northwest half of the Gulf (boxes 1–8, longitude west of 53°E). Salinity in boxes 9–13 (e.g., longitude 53°E to 55.5°E) decreases from 40–41 psu in January–February to 39.5–40.5 psu in March–June and to 39–40 psu in July–August. Density of deep water, how-

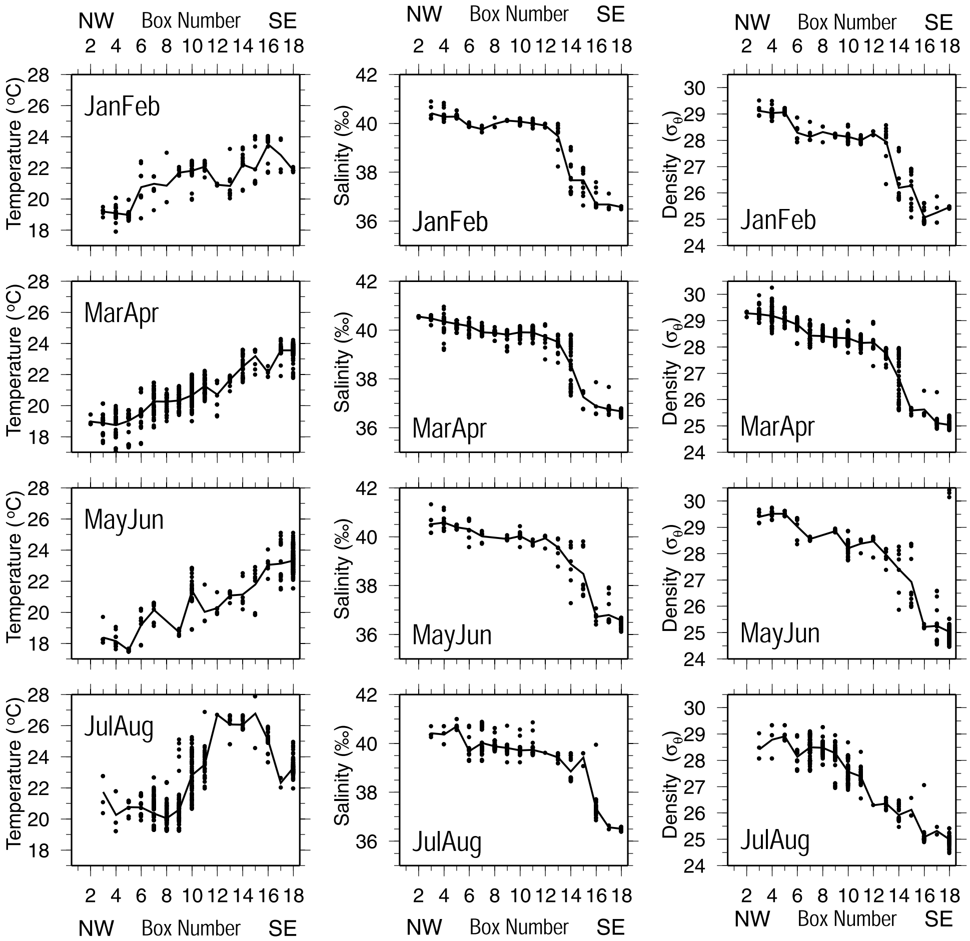

Figure 10. Axial sections at 50 m water depth (see Figure 2 for location of boxes) show little change from January to June. In July–August, deep water throughout Gulf warms and its density decreases. In the western approach to the Strait (boxes 12–15) this warming is much more dramatic and decreases the slope of the axial gradient in density. We attribute warming in this region to higher year-round rates of vertical mixing and a warming of sea surface water. The Strait is located at the boundary between boxes 15 and 16.

ever, changes significantly only from May–June to July–August, and this change is due to a warming of ~2°C of deep water, in the Gulf west of 54.5°E (boxes 1–11), and a warming of ~4°C in the Strait and its western approaches (boxes 12–16; longitude 54.5°E–57°E). The maps of salinity at 50 m depth (Figure 9) indicate only minor changes during the year.

[34] Seasonal changes in the contrast in density between the interior of the Gulf and the Gulf of Oman are too small to produce changes in outflow through the Strait. Figure 10 summarizes the seasonal trends along the Gulf axis at 50 m. Temporal and spatial patterns for deeper depth intervals are similar. In Figure 10 the along-axis salinity distribution remains nearly constant during the year. The temperature distribution remains constant from January to June but warms in July–August. This temperature increase is greatest close to the Strait (boxes 12–16), and, as a result, the density distribution flattens. Thus the density gradient from the western approaches through the Strait weakens significantly in summer. Since *Johns and Olson* [1998] observed that deep flow through the Strait is relatively constant throughout the year, deep flow through the Strait is unlikely to be driven by the local east-west density gradient near the Strait. There is, however, only a small change in the density difference between the central Gulf and the Gulf of Oman through the year. Density at 50 m remains σ_θ = ~25 in the Gulf of Oman (boxes 17–18), whereas the density at the other end of the Gulf remains σ_θ = 29–29.5 from January to June and decreases to σ_θ = 28.5–29 in July–August. This decrease amounts to a change of only 11–12.5%. A change of comparable magnitude in discharge out of the Gulf may not be resolved in the Johns and Olson mooring data because it could be accommodated by small changes in the thickness of the outflow or by changes in speed or cross-sectional area outside the channel axis. Thus deep flow out of the Gulf is likely driven by the density difference between deep water at the head of the gulf and near-bottom water on the Oman-Iranian shelf.

3.5. Vertical Mixing in the Western Approach to the Strait of Hormuz

[35] Unusually warm water temperatures throughout the water column characterize the western approach to the Strait of Hormuz in summer (July–August). In the axial

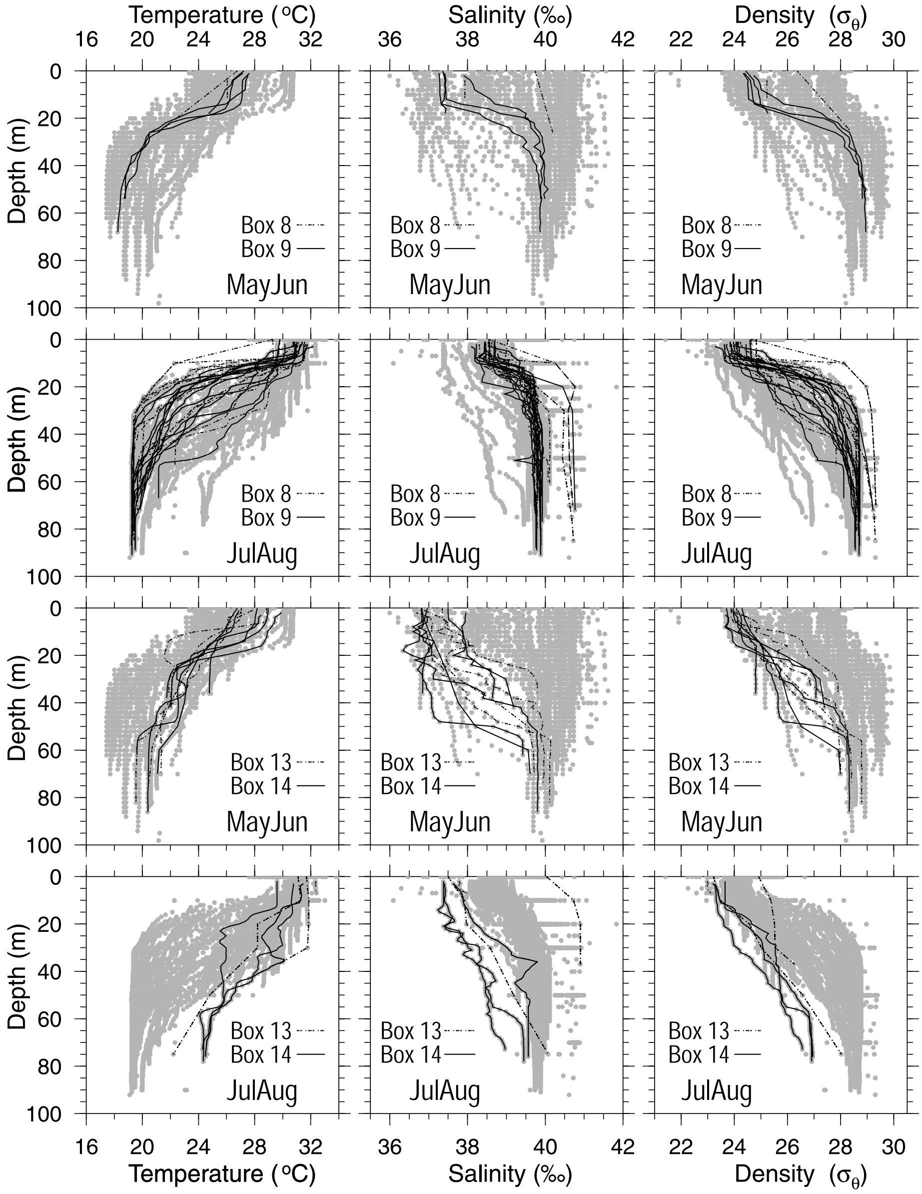

Figure 11. Casts made during the transition from spring to summer in boxes 8–9 (lines in top six panels) northwest of the high vertical mixing rate region are compared to casts in boxes 13–14 (lines in bottom six panels) inside the western approach to the Strait and to all casts in the Gulf (shaded dots). In both regions sea surface temperature and salinity increase between May–June and July–August reflecting solar warming and retreat of IOSW. Casts in boxes 8–9 differ little from mean Gulf profiles in both seasons, showing distinct surface mixed layers and steep seasonal thermoclines. In contrast, the warm surface temperature and low surface salinity in boxes 13–14 are mixed much deeper reaching to the seafloor in July–August. As a result, the density increase with depth is uniform without the characteristic summer layering elsewhere in the Gulf.

cross-sections through the western approach to the Strait (boxes 12–15, longitude 54.5°–56.5°E, Figure 2 shows box locations), warmer temperatures and lower densities reach closer to the seafloor in July–August than in May–June (Figures 7a, 7b, 7c, and 7d). Warmer than normal temperatures also appear in plots of properties at 50 m depth (Figure 10). Between June and July, the temperature of water at 50 m depth within the Strait and just to the west (boxes 12–16) changes from 0.5–2.5°C less than Gulf of Oman water (box 18) to 1.5–3.0°C warmer than Gulf of Oman water. Although salinity in boxes 12–16 appears to change little between June and July in Figure 10, density clearly decreases. Plots of property versus depth for the western approach (bottom two panels in Figure 11) show similar temporal changes in temperature and density at most depths and a decrease in salinity at water depths below 50 m.

[36] The unusual conditions are spatially restricted to the region extending westward from box 16 at the tip of the Musandam Peninsula to about box 12, bounded on the west

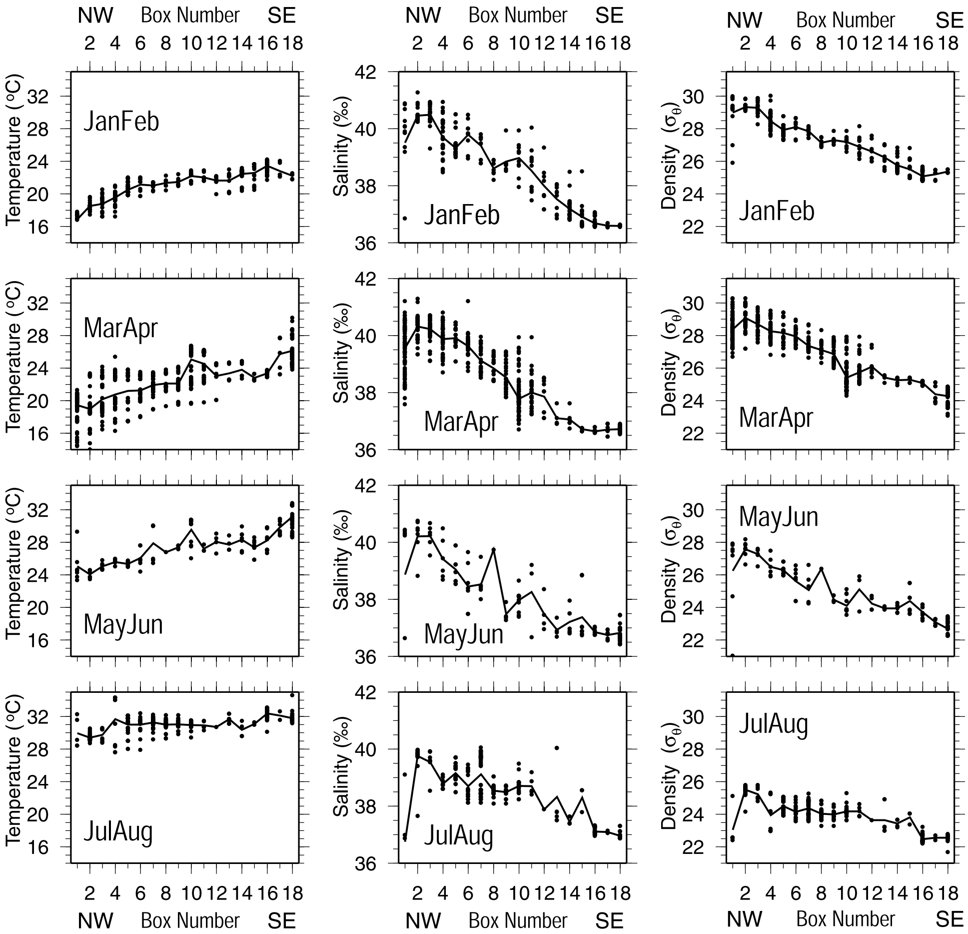

Figure 12. Axial sections at the sea surface (0 m water depth; see Figure 2 for location of boxes) show dramatic warming, especially in the late spring and summer. In July–August, sea surface temperature in the Gulf becomes nearly uniform with little difference across the Strait (boxes 15–16). In contrast, salinity changes little from January to June. In July–August, surface salinity decreases at the head of the Gulf (boxes 1–4) but increases elsewhere as low salinity IOSW retreats to the Strait. Throughout the year, density decreases toward the Strait reflecting the gradients in salinity and temperature. The gradient decreases only slightly in the summer. In winter, high density water ($\sigma_\theta > 29.5$) reaches the sea surface at the head of the Gulf.

by 54.5°E longitude. To the east of this region, little change occurs at 50 m depth in either temperature or salinity between June and July on the continental shelf between Oman-Iran (box 17) or in the Gulf of Oman (box 18; Figure 10). To the west of this region, water temperature at 50 m depth increases between June and July in box 11 (Figure 10), but salinity and density do not change significantly. Profiles of properties in boxes 8 and 9 show few similarities with profiles in the western approach (Figure 11). The western edge defined by water properties appears to occur near three islands aligned north-south along 54.5°E (Figure 1).

[37] Other investigators have found collaborative results. *Emery*'s [1956, Figure 5] plot of August temperature shows high temperatures extending to 45–55 m deeper water depths in this region than the central Gulf. *Johns and Olson* [1998] found that the vertical temperature profile observed at their mooring in the channel west of the Strait becomes much warmer in summer. These results suggest that our observations are not an artifact of the averaging done in our analysis.

[38] The appearance of unusually warm temperatures throughout the water column in the western approach to the Strait coincides with rapid warming of the sea surface. Warming of the sea surface affects the Gulf as a whole throughout the spring and summer. Figure 11 shows a warming at the sea surface between May–June and July–August of at least 3°C in axial boxes 13 and 14 and at least 4°C in boxes 8 and 9 in the central Gulf. Sea surface temperature increases of 3–6°C occur throughout the Gulf between month pairs during the spring (Figure 11). By July–August the sea surface in the Gulf has a nearly uniform temperature of 30–32°C, which is identical to that of the sea surface in the Gulf of Oman (Figure 12). Since the warming appears so uniform across the Gulf throughout the spring and summer and coincides with the development of a steep seasonal thermocline [*Reynolds*, 1993], we attribute the temperature increase to solar heating rather than to influx of warm surface water from the Gulf of Oman.

[39] We suggest that warm water temperatures observed at 50 m depth in the western approaches to the Strait (Figure 10)

are due to higher rates of vertical mixing there than elsewhere. As the surface water warms during the spring and summer, the higher temperatures are mixed downward to near-bottom depths. The difference in temperature between the surface and 50 m depth is 5–6°C in both the May–June and July–August periods indicating that the mixing rate does not change significantly with time. Thus the spike in temperatures at 50 m depth during July–August in Figure 10 is due to spatial variation in vertical mixing rather than temporal variation. Large, irregular vertical variations in temperature and salinity, as well as density inversions, are more common in the western approach to the Strait than elsewhere (Figure 11). Although direct measurements of vertical mixing have yet to be made, vertical mixing appears to be more effective in the western approach to the Strait of Hormuz than in the central Gulf to the north and higher than the rate on the Oman-Iran continental shelf and Gulf of Oman outside the Strait.

[40] The alternative hypothesis that there is a temporal change in vertical mixing suffers for lack of a feasible explanation. There are no changes in meteorological forcing, of which we are aware, that would cause vertical mixing of warm surface water to deeper depths to increase at the beginning of July. Although there is change in prevailing wind direction in the Strait itself between June and July, coincident with the strengthening of the summer monsoon, mean wind speed drops between the two months [*Brower et al.*, 1992]. The incidence of storms in the Gulf decreases at the beginning of July [*Perrone*, 1979]. New data in the future may reveal a source of forcing, but spatial variability in mixing appears to explain more observations at the present time.

[41] Another alternative hypothesis is that the warm temperatures at 50 m depth indicate the advection into the Strait of warm, salty water formed on the shallow banks to the south off the UAE. Unfortunately, there are no hydrographic transects across the southern edge of the Strait onto the UAE shelf in July–August (Figure 3), so direct data to confirm this interpretation is lacking. Indirectly, warm temperatures are consistent with this notion, but the maps of salinity at 50 m depth (Figure 9), however, indicate the salinity of most of the water at 50 m depth does not change significantly between May–June and July–August. Moreover, vertical profiles of salinity below 20 m depth (Figure 11) indicate that the salinity of water in the western approach is lower than salinity elsewhere in the basin rather than higher. Introduction of a new summer water mass from off the coasts flanking the Strait is unlikely because the higher temperatures we find are typically accompanied by normal or lower salinity, whereas high salinities, as well as high temperatures, are common in bays and on shallow banks of the Gulf in summer [*Sugden*, 1963; *Chandy et al.*, 1990]. New warm water from the coast should produce a characteristic salinity signal that we don't observe. High rates of vertical mixing accompanied by surface heating is a more likely explanation.

[42] We suggest that the reason for high rates of vertical mixing in the western approach to the Strait is interaction of tidally forced flow with seafloor topography. Whereas the Iran-Oman shelf to the east of the Strait is open and largely free of seafloor shallower than 50 m, the western approach to the Strait includes numerous islands, shoals, and irregular changes in channel depth. The northern edge of the approach is forced southward at least 18 km off the western tip of Qeshm Island by a broad, shallow flat formed by river-supplied sediment (Figure 1; see also Figures 1 and 14 of *Uchupi et al.* [1996]). Islands in the approach are perched on submarine pedestals 11–15 km across in a north-south direction that further constrict currents (Figures 1 and 2). Tidal flows near these islands are likely to be as accelerated by the topographic narrowing of the channel, as tides are around the tip of the Musandam Peninsula. We suggest that these accelerations are accompanied by formation of eddies and higher rates of vertical mixing that reach to the seafloor. There is geological evidence for this hypothesis. Seafloor depths reach 50–80 m deeper in channels that split to the north and south of some of these islands indicating that seafloor sediment is eroded by turbulent boundary layer currents. The only significant occurrence of gravel in the Gulf occurs in these channels and those that bend around the tip of the Musandam Peninsula [*Emery*, 1956; *Hartmann et al.*, 1971], indicating that scour driven by turbulent eddies is most effective there. West of the islands along 54.5°E, fewer topographic constrictions occur, so vertical mixing rates are lower.

4. Conclusions

[43] Seasonal and spatial variations in hydrographic data collected in the Persian/Arabian Gulf indicate features of water mass formation and circulation that have not been previously described.

1. The densest water forms during winter in shallow water at the northern end of the Gulf. The densest isopycnals outcrop there from January to April.

2. High salinity water forms along the western and southern Arabian coastlines, but the highest densities are not observed there because winter temperatures are milder than those in the north and considerable dilution occurs before high salinity water formed in shallow bays reaches the main Gulf basin.

3. A prominent pycnocline separates modified IOSW at the surface from Gulf Deep Water below. The density gradient is continuous with a front that separates IOSW from higher-salinity shelf water along the southern coast. The deeper isopycnals in the pycnocline intersect the seafloor in the Strait and upper isopycnals outcrop at the sea surface in the interior of the basin. Evidence suggests that there is little shear associated with the density gradient, and that Gulf Deep Water exiting the Gulf is modified by mixing across the density gradient.

4. The flow of modified IOSW into the Gulf, forming a low-salinity surface layer, peaks in May–June. Seasonal changes in the flux of IOSW may be driven by changes in sea surface slope caused by varying rates of evaporation.

5. We attribute the lack of significant seasonal variability in the flux through the Strait reported by *Johns and Olson* [1998] to relatively small seasonal changes in the density contrast between Gulf Deep Water and water at comparable depths on the Oman-Iran continental shelf outside the Strait.

6. Compared to the Gulf as a whole and the continental shelf outside the Gulf, the rate of vertical mixing is highest in the Strait of Hormuz channel around the Musandam Peninsula and in the 200 km long western approach to the

Strait. This distinction appears to be unrelated to meteorological forcing. We attribute the high rate to numerous topographic features that constrict the channel, accelerating tidal flows and generating large eddies capable of scouring the seafloor.

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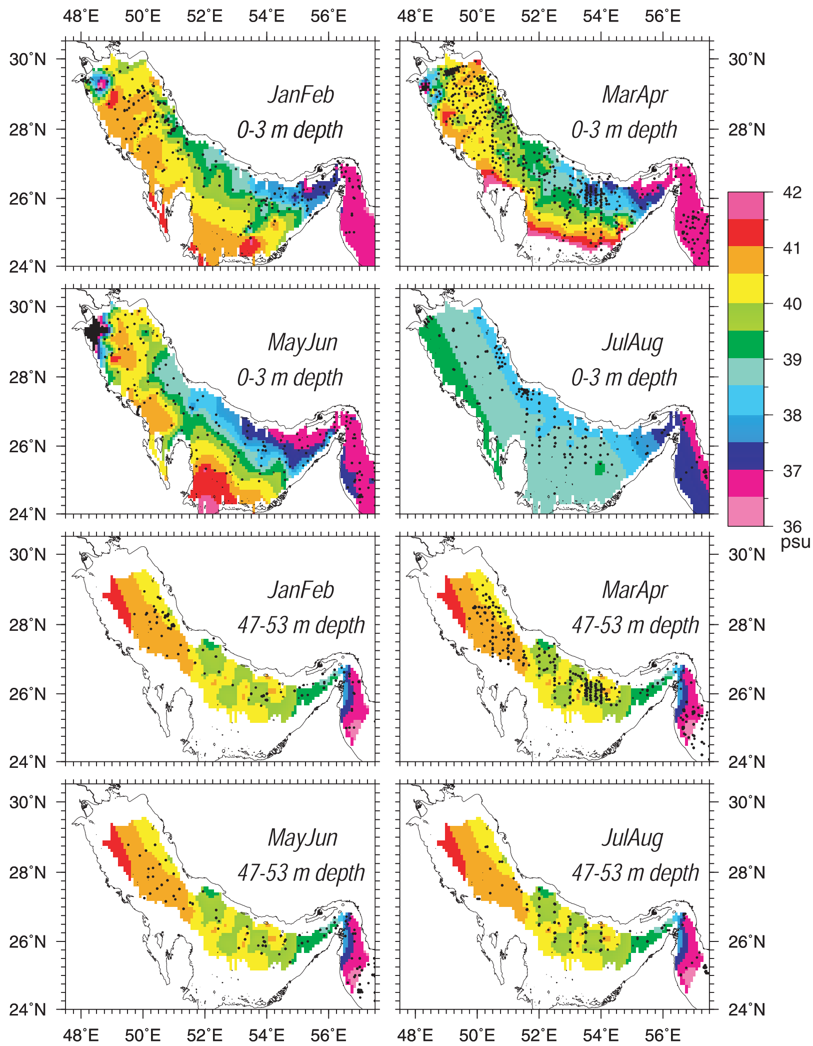

Figure 9. Surface salinity maps (top four panels) show that modified IOSW progressively moves farther up the Gulf from January to June. These data suggest that surface water with salinity less than 38 psu then retreats ~100 km eastward toward the Strait in July. Since the seasonal thermocline develops in May–June (Figures 5 and 7), this retreat is unrelated to summer warming. Salinity distribution of Gulf Deep Water at 50 m depth (bottom four panels) shows comparatively little seasonal change.