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Vertical, horizontal, and temporal changes in temperature in the Atlantis II and Discovery hot brine pools, Red Sea

Stephen A. Swift [a,*], Amy S. Bower [b], Raymond W. Schmitt [b]

[a] MS 24, Woods Hole Oceanographic Institution, Woods Hole, MA 02543, USA
[b] MS 21, Woods Hole Oceanographic Institution, Woods Hole, MA 02543, USA

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ABSTRACT

In October 2008, we measured temperature and salinity in hot, hypersaline brine filling the Atlantis II and Discovery Deeps on the Red Sea spreading center west of Jeddah, Saudi Arabia. In agreement with previous observations in the Atlantis II Deep, we found a stack of four convective layers with vertically uniform temperature profiles separated by thin interfaces with high vertical temperature gradients. Temperature in the thick lower convective layer in the Atlantis II Deep continued to slowly increase at 0.1 °C/year since the last observations in 1997. Previously published data show that the temperature of all four convective layers increased since the 1960s at the same rate, from which we infer that diffusive vertical heat flux between convective layers is rapid on time scales of 3–5 years and, thus, heat is lost from the brine pools to overlying Red Sea Deep Water. Heat budgets suggest that the heat flux from hydrothermal venting has decreased from 0.54 GW to 0.18 GW since 1966. A tow-yo survey found that temperature in the upper convective layers changes about 0.2 °C over 5–6 km but the temperature in the lower brine layer remains constant. Temperature in the lower convective layer in the Discovery Deep remains unchanged at 48 °C. To explain these results, we hypothesize that heat flux from a hydrothermal vent in the floor of the Discovery Deep has been stable for 40 years, whereas temperature of the brine in the Atlantis II Deep is adjusting to the change in hydrothermal heat flux from the vent in the Southwest Basin. We found no changes in the upper transition layer at 1900–1990 m depth that appeared between 1976 and 1992 and suggest that this layer originated from the seafloor elsewhere in the rift.

1. Introduction

In October 2008, we surveyed water mass properties in and around the Atlantis II and Discovery brine pools (Fig. 1). This paper describes our results and our effort to use these observations to extend the understanding of processes affecting the brine pools determined by previous surveys.

The Atlantis II and Discovery Deeps in the Red Sea were the first hydrothermal sites discovered in the oceans (temperatures of about 56 °C and 45 °C, respectively, in 1965), and their brine pools have been monitored the longest. Anomalously high temperatures were serendipitously observed near the seafloor along the axis of the Red Sea in 1948 by the *R/V Albatross* and in 1958 by the *R/V Atlantis*. The results, however, were not published in the open literature (Bruneau et al., 1953; Neuman and Densmore, 1959; Miller, 1969; Swallow, 1969). The anomaly was confirmed in 1963 by sampling from the *R/V Atlantis II* and *R/V Discovery* as they transited through the Red Sea during the International Indian Ocean Expedition (Miller, 1964; Swallow and Crease, 1965). The hot, saline brines of the Atlantis II and Discovery and Chain Deeps were first systematically investigated and mapped by the *R/V Chain* in 1966 and the results presented in Degens and Ross (1969). Munns et al. (1967) reported an increase of 0.56 °C between February 1965 and October 1966. Brewer et al. (1971) showed that temperatures in two vertically uniform brine layers of the Atlantis II Deep had increased by 2.7 °C (lower convective layer, LCL) and 5.6 °C (upper convective layer, UCL1) from November 1966 to February 1969. Subsequent measurements showed that temperatures continued to increase reaching 67.1 °C in 1997, although the rate of increase slowed somewhat during the 1970s (Bubnov et al., 1977; Schoell and Hartmann, 1978; Hartmann, 1980; Monin and Plakhin, 1982; Blanc and Anschutz, 1995; Hartmann et al., 1998a, 1999b). In contrast, the temperature of the lower layer in the Discovery Deep remained constant at about 44.7 °C.

The vertical temperature structure of the brine in the two basins differs, as well. Munns et al. (1967) showed that brine in

* Corresponding author. Tel.: +1 508 289 2626; fax: +1 508 457 2150.
E-mail addresses: sswift@whoi.edu (S.A. Swift), abower@whoi.edu (A.S. Bower), rschmitt@whoi.edu (R.W. Schmitt).

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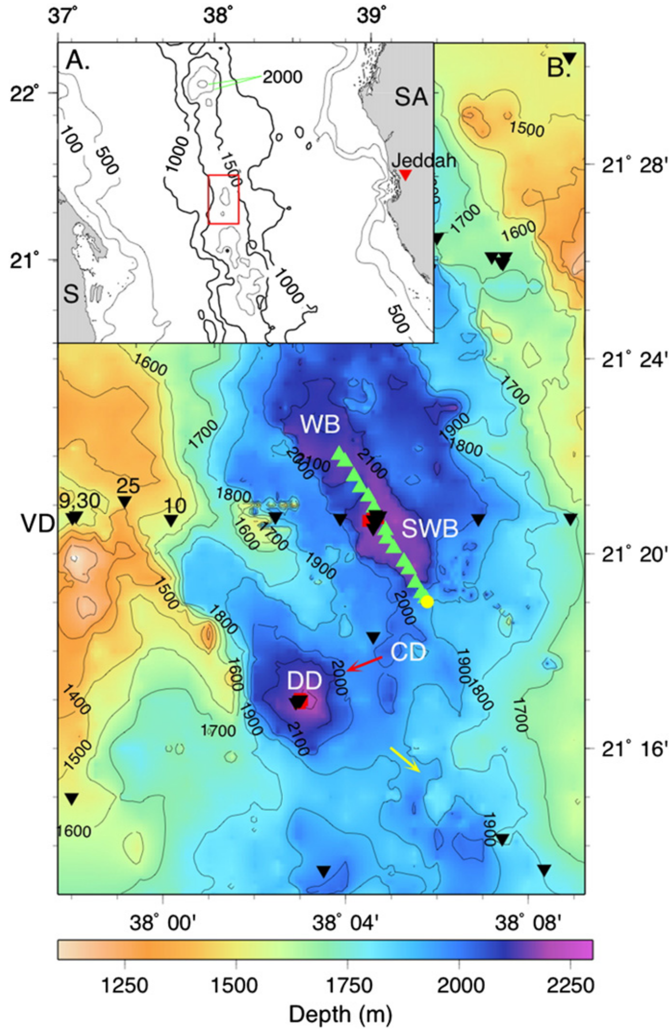

Fig. 1. (a) Index map shows location of *R/V Oceanus* Cruise 449-6 survey area (red box) in the Red Sea between Saudi Arabia (SA) and Sudan (S). Bathymetry is derived from satellite gravity (SRTM30, D. Sandwell). (b) Bathymetry map compiled from a grid of single-beam echosounding lines (Bower, 2009) shows location of hydrographic casts with Hobo temperature sensors (red squares), both Seabird and high-range CTDs (black inverted triangles), and the locations of the deep turning points (green triangles) during the Hobo tow-yo. Hydrographic data were collected in the Southwest Basin of the Atlantis II Deep (SWB), Discovery Deep (DD), Chain Deep (CD), and the Valdivia Deep (VD). The tow-yo extended northwestward from the Southwest Basin into the West Basin (WB). Yellow dot marks the location of the origin for the range scale in Figs. 7 and 8. Red arrow marks the sill at 1990 m between the Chain basins and the Discovery Basin. Yellow arrow marks the spillover depth at 1902 m for the brine pool complex. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

the Atlantis II Deep in 1965 comprises (1) a high-temperature (56.5 °C) bottom layer below about 2050 m depth with thickness greater than 140 m and (2) a 100 m thick transition layer characterized mid-way by a 30–40 m thick layer with uniform temperature (44 °C). From continuous temperature measurements obtained in 1992, Blanc and Anschutz (1995) recognized three uniform layers in the transition layer (UCL1–3) separated by thin interfaces with thicknesses of 1–2 m. Although early observations in the Discovery Deep by Ross and Hunt (1967) also found a single uniform layer at about 36 °C interrupting the vertical temperature gradient in the transition layer, subsequent investigations found a continuously varying transition layer (Bubnov et al., 1977; Danielsson et al., 1980; Winckler et al., 2001; Schmidt et al., 2003). Thus, the transition layer in the Atlantis II basin became more structured, whereas the transition layer in the Discovery Basin became less structured.

Another temporal change observed since the 1960s is the growth in thickness of the transition layer of both basins. Early measurements placed the base of Red Sea Deep Water above both basins at about 1945 m depth (Swallow and Crease, 1965; Brewer et al., 1965; Munns et al., 1967; Ross, 1972; Bubnov et al., 1977). In 1992 Blanc and Anschutz (1995) observed a second transition layer with a distinctly lower vertical temperature gradient that extended up to about 1900 m depth. Observations reported by Winckler et al., 2001 and Schmidt et al. (2003) indicate that this new transition layer was present in 1997. In calculating heat and salt fluxes for the deep, Anschutz and Blanc (1996) treated the new layer as an expansion of the transition zone due to addition of new heat from the seafloor.

Spatial as well as vertical variations in the temperature of the lower layer were observed in 1971. Schoell and Hartmann (1973) found a drop of about 1.7 °C in maximum temperature of the lower brine layer over >12 km from the Southwest Basin of the Atlantis II Deep to the North Basin. To explain these observations, they hypothesized a hydrothermal vent in the Southwest Basin and a clockwise "spreading" of recently injected hot water through the sub-basins of the Atlantis II brine pool. In 1980, Monin and Plakhin (1982) observed a similar spatial change in temperature of the LCL and reported a northward decrease in temperature of the UCL1, as well. The gradient in the lower brine layer, however, is not a permanent feature of the basin. Blanc and Anschutz (1995) re-sampled the basins in 1992 and found no significant lateral temperature differences.

Previous investigators concluded that the vertical structure of brine fluids is consistent with processes of turbulent convective mixing in the uniform layers and double diffusion across the interfaces between the convecting layers (e.g., Turner, 1969; Schoell and Hartmann, 1973; McDougall, 1984; Blanc and Anschutz, 1995). Double diffusion is characterized by faster molecular diffusion of heat than salt with opposing effects on density (Schmitt, 1994). Turner (1969) first applied this model to the Atlantis II Deep based on similarities of early field observations to the results of theoretical and laboratory models of double diffusion. He suggested that the addition of heat and salt from below occurs in a turbulent plume, presumably from a hydrothermal vent, although no such vent has yet been directly observed in the Atlantis II Deep. Turner further supposed that vent turbulence established the lower convective layer, whereas the intermediate uniform layers originated by mixing accompanying the breaking of internal waves along the sides of the basin. In the model, convection in the intermediate layers is maintained by more rapid diffusion of heat than salt across their lower boundaries. Later field observations confirmed several features of Turner's hypothesis. Internal waves on a brine surface were observed by submersible (Monin and Plakhin, 1982). The system of mixed layers and high gradient interfaces also resembles the laboratory experiments of Huppert and Linden (1979) that involved heating a stable salinity gradient from below. Application of a heat flux caused layering to propagate upward into the salt gradient, with lower layers growing by merging and diffusive fluxes across the interfaces driving thermal convection in the mixed layers. In the Atlantis II Deep, Voorhis and Dorson (1975) placed an instrument that recorded vertical current flow in the upper convecting layer, and recovered the current meter 3.5 day later after a lateral drift of 3 km to the east for a mean drift of about 1 cm/s. The results showed turbulent vertical motion on scales of 1 m or less and unsteady convection on the scale of the layer (30 m).

Erickson and Simmons (1969) measured sediment temperature in both the Atlantis II and Discovery Deeps. In some cores, they found non-linear increases in temperature, whereas in others, they found no change in temperature with depth below seafloor. The maximum temperature was 62.5 °C. Assuming that some cores over-penetrated, they linearized temperature gradients just below the seafloor and computed conductive heat flow of 630–840 mW/m^2 for the Atlantis II Deep and 500–630 mW/m^2 for the edge of the Discovery Deep.

There have been no measurements of the hydrographic properties of the Atlantis II or Discovery Deeps since a visit by the *R/V Sonne* in 1997 (Hartmann et al., 1998b; Winckler et al., 2001; Schmidt et al., 2003). In October 2008, the *R/V Oceanus* visited the Red Sea to make hydrographic and microbiological observations. During this cruise, a number of opportunistic hydrocasts were made into the brine layers to determine whether changes in temperature and salinity had occurred during the past decade.

2. Methods

Most of the new data collected from the Atlantis II and Discovery Deep brine pools were obtained using one of two internally-recording temperature sensors manufactured by Onset Computer Corporation of Onset, Massachusetts. The "Hobo" Stainless Temperature Data Loggers with 5-inch probes, Model U12-015-02, were set to sample at 1 Hz. PVC housings were manufactured to provide some protection for the external probes during deployment and recovery. The Hobo logger was attached either to the hydrowire or to the handle of a Niskin bottle on the hydrowire and lowered to near the sea floor using a pinger to measure distance off the seafloor.

The Hobo manufacturer's specifications indicate an accuracy of ± 0.2 °C at 25 °C, a resolution of 0.025 °C and a response time of 20 s to reach 90% of value. Bench testing indicated that the sensors performed better than specifications. Prior to the cruise, both instruments were calibrated (Fig. 2) and pressure tested to 2500 dbar using facilities at the Woods Hole Oceanographic Institution. The r.m.s. difference between the sensors and the standard was 0.07 °C and 0.02 °C for the two probes in the temperature range 30°–60 °C against an ASL-F18 resistance bridge with a standard platinum thermometer. Total error of the calibration temperature measurements from all sources is ± 0.020 °C. Two years after the cruise, re-testing showed drift of <0.04 °C at temperatures $<=60°$ (Fig. 2). At 70 °C instrument S/N 2005482 differed from the standard by -0.076 °C. Assuming

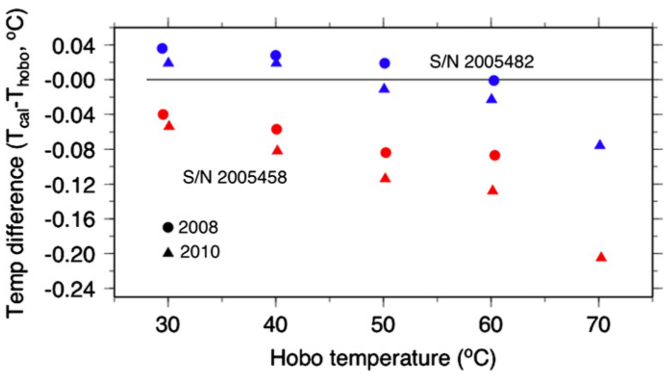

Fig. 2. The two Hobo sensors were calibrated immediately before (2008) and 2 years after (2010) the cruise in a stirred constant temperature bath. During calibration, each Hobo was totally (probe plus electronics) immersed in the bath. Calibration temperature was measured to ± 0.02 °C with an ASL-F18 resistance bridge with a standard platinum thermometer. The data indicate drift over two years of <0.04 °C at 60 °C or less. All Hobo temperatures are averages of 300 consecutive 1-s data points. Hobo output was stable within 0.005 °C for at least 15 min prior to start of observations.

the drift at 60° ($-0.022°$) applies at 70° and that all the drift occurred after the cruise, the error is reduced to -0.054 °C. For instrument S/N 2005458, the drift at 60 °C ($-0.041°$) reduces the error to -0.164 °C. Thus, both instruments were accurate to better than ± 0.16 °C at the time of the cruise. By plunging the sensors in an ice bath, we estimated response times to be about 8 s to reach 90% of value. A time lag correction was made to the Hobo temperatures using a single-pole filter with a time lag of 2.5 s. This time lag produced the most homogenous brine layers without overshooting at the brine interfaces.

The lack of a pressure transducer on the Hobo meant that depth estimates could only be made using the ship's digital log of wire out and in some cases, the distance of the acoustic pinger from the sea floor. To improve accuracy, we calibrated Hobo depth to CTD depth using temperature. The SeaBird CTD temperature measurements in the transition layers over the Atlantis II and Discovery Deeps showed a consistent depth (within ± 1 m) of the 28 °C isotherm: 1996.5 m for Atlantis II and 2022.5 m for Discovery. Depths of all Hobo profiles were adjusted to match the depth of the 28 °C isotherm.

To obtain a vertical cross section of temperature across the Atlantis II brine pool, we towed a Hobo near the end of the hydrowire while repetitively raising and lowering the cable. The "tow-yo" proceeded northwestward at about 1.3 knots from the Southwest Basin into the West Basin (Fig. 1). The shallowest depth reached during the tow was $\sim$1700 m. The turns at the seafloor are spaced $\sim$0.5 km apart.

Two temperature and salinity profiles were also obtained in the Southwest Basin of the Atlantis II Deep using a custom-built high-range CTD developed at Woods Hole Oceanographic Institution. The instrument is an internally-recording CTD built into a titanium housing with special sensors which are expected to give results up to 6500 m depth, 70 °C and 500 ppt salinity, and the ability to provide its temperature signal to the SBE9 CTD for real-time reading when suspended from the Rosette on a 50 m cable. The high-range CTD thermometer was calibrated to 2 millidegree accuracy in the range of 10–70 °C. The conductivity cell was calibrated in the normal oceanographic range as no high conductivity standards are available. Also, since no alternative formula is available, salinity values were computed from conductivity by applying the equation of state for normal seawater, though it is not expected to be accurate at these high temperatures and salinities, nor is there any assurance that the ratio of constituents is the same. Since we did not measure salinities on brine water samples, no corrections to true salinity have been applied. Nevertheless, vertical changes of our values relative to temperature are useful to understanding the processes controlling the vertical structure (see Section 4.1). In addition, a number of profiles of the transition layer between Red Sea Deep Water (RSDW) and the convective layers were obtained with a standard Seabird SBE 911 CTD. These data extend down to brine temperatures of about 32 °C, which is just below the upper limit of the SeaBird instrument. Using pre-cruise calibration data and instrument drift rates, the temperature measurements are accurate to ± 0.0005 °C. After standard conductivity processing, salinity data were adjusted to fit values measured on bottle samples at-sea using a titration salinometer accurate to ± 0.003 psu. The corrected salinities have a nominal error of about ± 0.01 psu relative to standard seawater. Winkler titration measurements for oxygen in bottle water samples have a mean error of ± 0.02 ml/l (Knapp et al., 1990).

Water depths were recorded at three second intervals ($\sim$15 m at 10 knots) by a hull-mounted Knudsen chirp echosounder at $\sim$12 kHz (Bower, 2009). Data were collected on a grid of north-south and east-west lines spaced about 1.8 km apart plus along transit lines between stations. Using results from the initial

CTD station, depths were corrected for variations in sound speed above the brine layer surface but not below. The shapes and locations of topographic features in our gridded database (Fig. 1) closely reproduce those in more detailed databases (e.g., Pautot, 1983). Sill depths between basins based on echosounding maps are taken from Anschutz and Blanc (1996) and Hartmann et al. (1998a): 2000 m between Atlantis II Deep and the Chain A Deep; 1990 m between Chain A Deep and Discovery Deep; and 1902 m for the sill emptying the Discovery Basin (and, thus, the entire brine pool complex) to the south.

We compute density ratio as $R_\rho = \beta\Delta S/\alpha\Delta T$ using Matlab routines by N.L. Bindoff (CSIRO) to obtain α, the coefficient of thermal expansion, and β, the coefficient of haline contraction using the normal equation of state. ΔS and ΔT were computed over 10 m depth to reduce noise from division by small differences. Density ratio is the "ratio of the separate contributions of salinity and temperature to the density difference between layers" (Turner, 1969, p. 166). Laboratory experiments indicate that molecular diffusion becomes more important than turbulent mixing as values of density ratio increase above about 7 (Turner, 1969; Turner, 1973, chapter 8).

3. Results – temperature and salinity structure

Fig. 3 shows the five Hobo temperature profiles from the center of the Southwest Basin of Atlantis II Deep (location in Fig. 1). The 2008 observations show that the four convective layers documented by Blanc and Anschutz (1995) and Hartmann et al. (1998a, 1999b) still exist, including the Lower Convective Layer (LCL) and Upper Convective Layers UCL1, UCL2 and UCL3. A

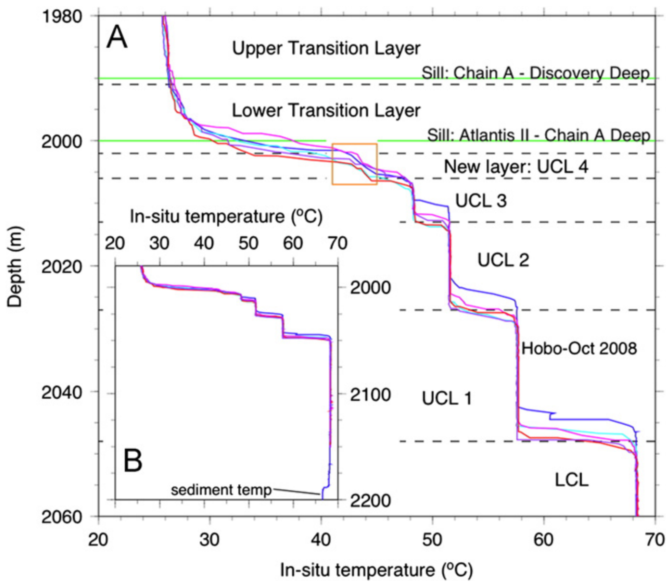

Fig. 3. (a) Uniform convective layers (LCL, UCL1–4) separated by thin (<2 m) double diffusive layers characterize the high gradient portion of temperature profiles in the Southwest basin of the Atlantis II brine pool. (b) In the entire brine pool profile, the thickness of the LCL is 135 m, and the temperature of the sediment is about 1.7 °C cooler than in the brine. All profiles were obtained with a Hobo thermometer with an accuracy of better than ±0.16 °C. Cast locations in Fig. 1. The orange square highlights a convective layer (UCL4) not recognized by previous investigations. Temperature increases downward gradually in the Upper Transition Layer (1900–1992 m) and rapidly in the Lower Transition Layer (1992–2002 m). Horizontal dashed lines mark depth boundaries of convective and transition layers; green lines mark sill depths taken from bathymetry by Anschutz and Blanc (1996) and Hartmann et al. (1998a). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

new convective layer UCL4 is developing above UCL3 (~42.5 °C, 2002–2006 m; orange box in Fig. 3(a)). Above UCL4, temperature decreases upward by about 16.5 °C in a thin (12 m) Lower Transition Layer (1990–2002 m depth). Above 1990 m depth, temperature continues to decrease but at a much lower rate forming an Upper Transition Layer (Fig. 4(a)). Blanc and Anschutz (1995) identified the same structure for the transition between the convective layers and RSDW. Salinity co-varies with temperature (Fig. 4(a)). Density ratio is only slightly elevated in the Upper Transition Layer but increases rapidly across boundaries between convective layers (Fig. 4(a)). The two-tiered structure of UCL2 observed by Blanc and Anschutz (1995) in 1992 was not detected in 2008. Temperature in the LCL is vertically homogeneous. At the seafloor, temperature cools by ~1.7 °C (Fig. 3(b)). Table 1 summarizes the October 2008 observations for all layers and interfaces.

In the Discovery Deep, we also found Upper and Lower Transition Layers: temperature below RSDW increases slowly from 22 °C at 1900 m to 26 °C at 2010 m and then rapidly up to 45 °C at 2058 m depth (Table 1; Figs. 4(b) and 5). In the LCL, temperature remains uniform down to 2132–2145 m where a small increase of ~0.16 °C occurs. In the Lower Transition Layer, temperature increases downward smoothly with minor changes in gradient on depth scales of 1–2 m. Unlike the Atlantis II Deep, no intermediate convective layers occur. Also unlike the Atlantis II Deep, density ratio begins to increase rapidly at ~1978 m, within the Upper Transition Layer (Fig. 4(b)). This depth coincides with the depth at which the temperature profiles in the two basins separate (Fig. 5; arrows in Fig. 4) and marks the effective sill depth for lateral water exchange between the two basins. Peak density ratio occurs at 1990 m, which is above the Lower Transition Layer as well.

Figs. 4 and 5 show that below the effective sill depth separating the two basins, the brines in both the transition layers and convective layers in the Atlantis II Deep are warmer than the brines at similar depths in the Discovery Deep. The separation in temperature occurs ~20 m above the Chain-Discovery Deep sill identified in bathymetric data. The middle panels in Fig. 4 show that salinity in the Lower Transition Layer is higher in the Atlantis II Deep than in the same layer in the Discovery Deep. The temperature-salinity plot in Fig. 6 indicates that the temperature at a given salinity below the effective sill depth in the Atlantis II Deep is higher than that in the Discovery Deep and, therefore, the density of a portion of the Upper and Lower Transition Layers in the Atlantis II Deep is lower than the density in the Discovery Deep. Fig. 6 also shows that this difference in density existed in 1976 and, thus, that the apparent difference in density is not an artifact of the way we computed salinity from conductivity. These results are consistent with similar *T–S* plots in Brewer et al., 1969 and Monin and Plakhin (1982). Thus, the difference in brine density between the two basins has been a persistent feature for the last 50 years of observations.

The deep tow-yo of the Hobo sensor in the Atlantis II Deep revealed lateral changes in temperature of the upper convective layers and lateral uniformity in the LCL. Fig. 7 shows the temperature of the various layers along the tow-yo track (location in Fig. 1) compared with the temperature at 1950 m depth at the base of the Upper Transition Layer (UTL). The temperature of the LCL layer remains essentially unchanged over 7 km. In the shallower convective layers, though, temperature changes little for the first 1–2 km and then decreases by about 0.2 °C over 5–6 km. The stability of the LCL temperatures indicates that this decrease is not due to instrument drift. In contrast, the temperature of the UTL increases by about 0.2 °C.

Previous studies described similar lateral temperature changes, but some of these gradients have changed since the

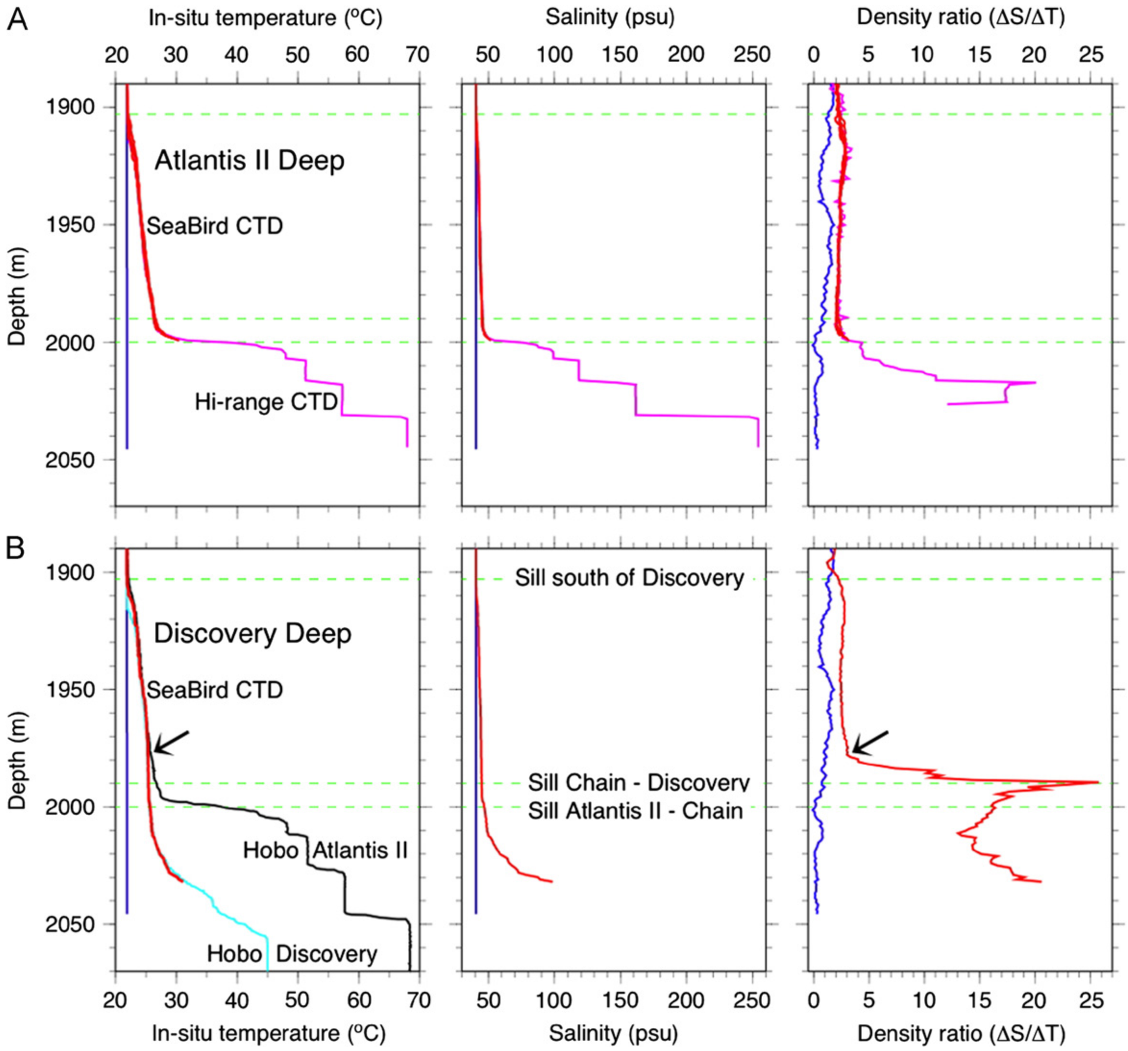

Fig. 4. Density ratio ($\beta\Delta S/\alpha\Delta T$; right panel) is computed for (a) Atlantis II Deep and (b) Discovery Deep and compared to profiles of in-situ temperature (left panel) and salinity (middle panel). Dashed green lines mark depths of sills based on bathymetry. A CTD station outside the brines (blue line) is included for comparison. In panels for the Atlantis II Deep (a), we show SeaBird CTD profiles (red) and High-Range CTD (magenta). No Hi-range CTD data were collected in the Discovery Deep, so we include a Hobo temperature profile (light blue) to define the depth to the LCL. In both basins, density ratio is slightly above background level in the Upper Transition Layer between 1902 m and 1990 m. In the Atlantis II Deep the significant increase in density ratio begins at the 2000 m sill depth coincident with sharp increases in temperature and salinity. In contrast, the significant increase in density ratio in the Discovery Deep occurs within the low gradient depth interval coincident with the separation of temperature profiles for the two basins (arrows; see also Fig. 5). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Table 1
Temperatures, salinity and depths of brine pool layers during October 2008.

	Southwest Basin-Atlantis II Deep			Discovery Deep	
	Temperature (°C)	Salinity* (psu)	Interface depth (m)	Temperature (°C)	Interface depth (m)
LCL	68.28	252	2048	44.99	2059
UCL1	57.32	162	2027		
UCL2	51.25	118	2013		
UCL3	47.98	99	2006		
UCL4	43.53	87	2002		

* Salinity was computed from conductivity using the equation of state for normal seawater.

1960s. In Fig. 8 we project some of these trends onto the track of our tow-yo. Schoell and Hartmann (1973) describe a clockwise reduction in maximum temperature of the LCL approaching 2 °C between basins of the Atlantis II Deep and local variations of 0.3 °C over ~2.5 km in the Southwest Basin. Their data collected in March-April of 1971 also include a northward decrease of ~0.1 °C in the temperature of UCL1 (Fig. 8(b)). This gradient is almost certainly real, because Voorhis and Dorson (1975) reported that the temperature in UCL1 in February 1971 increased by 0.1 °C southeastward along a drift of 3 km. Furthermore, they state that data from reversing thermometers corroborate the temperature gradient. In 1980 Monin and Plakhin (1982) found similar gradients and similar variability in the LCL (Fig. 8(a)). In the UCL1, however, Monin and Plakhin found that the temperature of two sublayers decreased by 0.2–0.3 °C over 4 km indicating gradients higher than in 1971 but similar to those that we found in 2008 (Fig. 7). By 1992 spatial differences in the temperature of the LCL between sub-basins had disappeared (Blanc and Anschutz, 1995), and by 2008 the temperature of the LCL in the southwest basin was uniform.

Our data from October 2008 extend temporal trends observed by previous sampling. Fig. 9 shows the temperature history of each convective layer back to the first systematic observations in the mid-1960s. Temperature in three of four layers in the Atlantis II Deep increased since *R/V Sonne* 121 visited in July 1997 (Fig. 9;

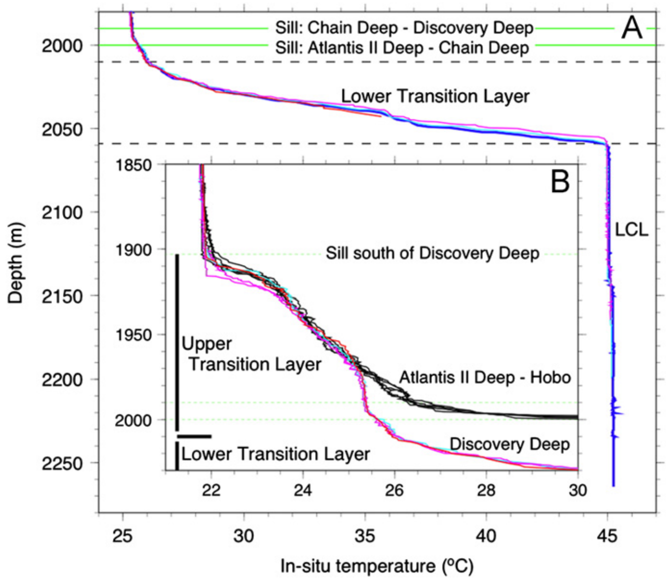

Fig. 5. In the Discovery Deep, (a) temperature increases in a near linear gradient from 2010 m down to a "lower" convective layer (LCL) below 2059 m. Minor wiggles may mark thin convective layers. Red line is a SeaBird CTD profile; all other lines are Hobo profiles. (b) In the Upper Transition Layer, temperature profiles above 1970 m depth in the Discovery Deep (colors) match profiles in the Atlantis II Deep (black) indicating a similar origin. Between 1970 m and ~2015 m, two thin (10–20 m thick) temperature gradients connect the low gradient layer downward to the top of the steep temperature gradient that forms the Lower Transition Layer. Sill depths (green lines) are based on bathymetry. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

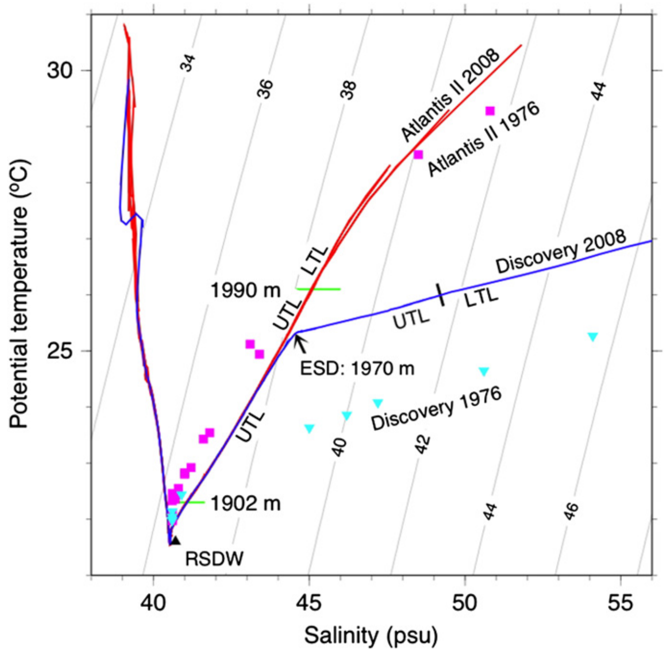

Fig. 6. A temperature-salinity plot with density increasing towards the lower right indicates that the Lower Transition Layer (LTL) in the Atlantis II Deep is warmer and, therefore, less dense than LTL brine in the Discovery Deep. Water depths apply to 2008 profiles in the Atlantis II Basin only. We computed 2008 salinity values >42 by extrapolation of the conductivity/salinity relationship for normal seawater. Corrections for brine chemistry would apply equally to both Atlantis II and Discovery brines and, thus, are unlikely to affect the relative density of the brines shown in the figure. Seabird CTD data from 2008 and bottle data from 1976 indicate that the density relationship has persisted for 30 years (Atlantis II Deep: red line [2008] and magenta squares [1976]; Discovery Deep: blue line [2008] and light blue inverted triangles [1976]). Bottle data are from Danielsson et al. (1980) and Bubnov et al. (1977). RSDW (black triangle) is from Sofianos and Johns (2007). Green lines mark present temperatures at bathymetric sill depths in the Atlantis II Deep. Thin black lines mark potential density computed for standard seawater at 1900 db. The effective sill depth (ESD) for flow between the two deeps occurs at ~1970 m depth within the Upper Transition Layer (UTL). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Hartmann et al., 1998b, Winckler et al., 2001). Temperature in the LCL increased by 1.08 °C over 11.25 years, for an average rate of 0.10 °C/year. This is the slowest rate of long-term temperature increase to date: it is less than 1/6 of the warming rate from 1966 to 1972 (0.62 °C/year) and less than 1/3 of the rate from 1972 to 1992 (0.30 °C/year) as reported by Blanc and Anschutz (1995). Fig. 9 shows that the rate of temperature increase has steadily decreased with time in all layers.

The slow-down in the warming rate of the LCL can be expressed in terms of a change in the rate of heating and compared to previous estimates. Anschutz et al. (1998) found that the heat content of the LCL had increased at a rate of 0.18 GW between 1977 and 1992. Using parameters from their Table 1 and temperature measurements in 1997 by Hartmann et al. (1998b), we obtained rates of 0.15 GW for 1992–1997 and 0.046 for 1997–2008. So, the rate at which heat is being added to the LCL has decreased. The rate for the most recent period is four times lower than the rate spanning the 1980s. In the next section, we suggest that this is associated with a slow-down in the rate of heat flux associated with hydrothermal venting.

4. Discussion of spatial and temporal changes

4.1. Convective layers

The rate of change in temperature for all four convective layers in the Atlantis II Deep is almost uniform since the early 1970s. This is particularly remarkable for time spans as short as 5 years between 1992 (Blanc and Anschutz, 1995) and 1997 (Hartmann et al., 1998b). This observation reveals an important aspect about the heat balance of the brine: upward diffusion of heat across the thin interfaces with high temperature gradients is rapid on the scale of years. If the exchange of heat between the LCL and RSDW was slowed on time scales greater than 3–5 years by any one convective layer and its diffusive lid, then the time rate of change in temperature in the two layers would differ. The parallel change in temperature history curves in Fig. 9 shows that lags are minor on these time scales. We infer that the convective layers and their bounding gradient layers provide no delay in vertical heat transport on time scales of 3–5 years. Thus, the temperature of the system is controlled by the flux and temperature of hydrothermal fluids entering the LCL through the seafloor. Heat gained by the brines from the hydrothermal vent is lost to the overlying RSDW, whose slow southward circulation (Cember, 1988; Sofianos and Johns, 2007) carries the heat away.

To estimate the hydrothermal heat flux, we compare the rate of heat gain for the LCL since the last temperature measurement in 1997 with the diffusive heat flow out the top of the LCL. Above, we found a net heat flux into the LCL of 0.046 GW between 1997 and 2008. This change must be balanced by heat exchange through the boundaries of the layer. The temperature of the LCL throughout this period exceeded sediment temperatures at all depths below seafloor measured by Erickson and Simmons (1969) in 1966, so the seafloor may be a heat sink. However, no subseafloor temperature gradients have been recently measured from which conductive heat flow could be confidently estimated. Using the approach of Anschutz et al. (1999, p. 1789), we estimate diffusive heat loss through the lid of the LCL to be 2.5 W/m^2 or 0.13 GW using a surface area of 51.7 km^2 (Anschutz and Blanc,

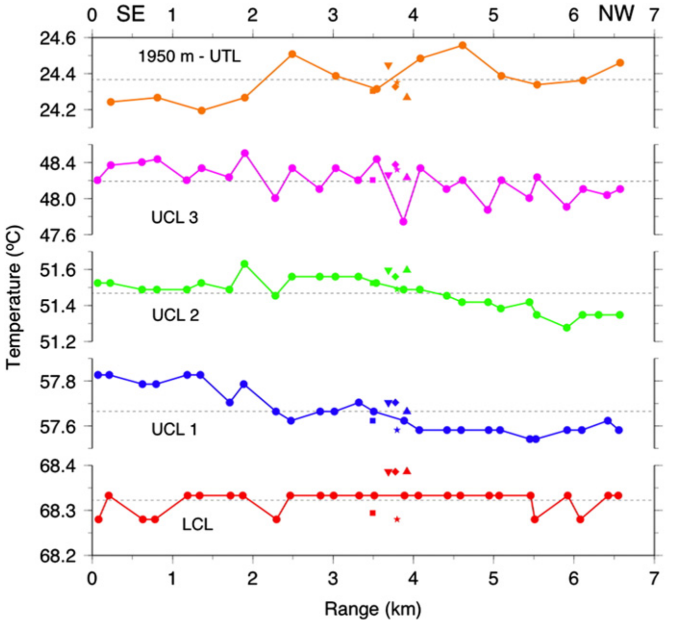

Fig. 7. Temperatures in convective layers UCL1–3 decrease by about 0.2 °C over 5–7 km from the Southwest Basin (left) to the West Basin (right) of the Atlantis II Deep. Fig. 1 shows the location of the Hobo tow-yo (green triangles) and the location of zero range (yellow dot). Over a similar distance, the temperature in the middle of the Upper Transition Layer (1990 m) increased by about 0.3 °C, and the temperature of the underlying LCL remained essentially unchanged. Unconnected points are independent measurements of temperature from CTD casts in the Southwest Basin (see Fig. 1 for location). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

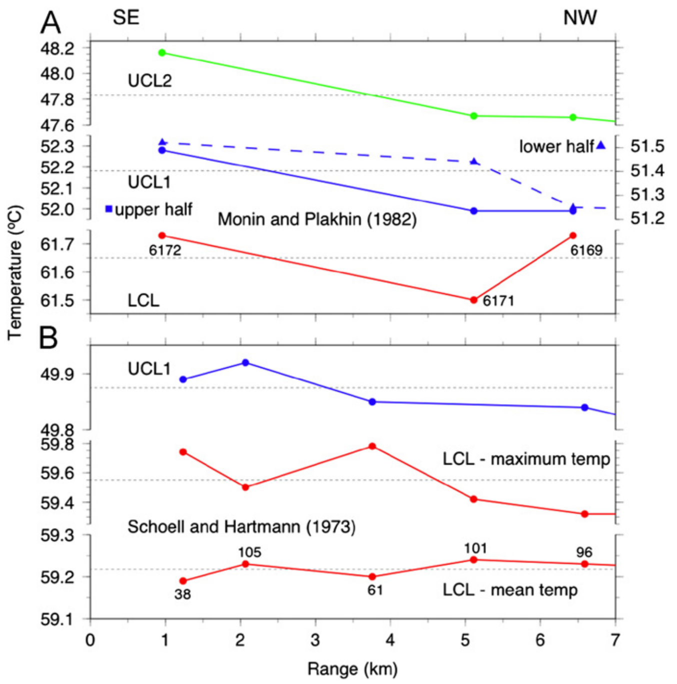

Fig. 8. The temperature of upper convective layers in data tabled by (a) Monin and Plakhin (1982) and by (b) Schoell and Hartmann (1973) decreases northward in the same direction as our tow-yo. Range origin is the same as in Fig. 7 (yellow dot in Fig. 1). Casts are numbered in plots for the LCL. Dashed lines mark mean values. The temperature decreases in the upper convective layers are greater (0.5 °C) in Monin's study but less (~0.07 °C) in Schoell's. A significant difference with spatial trend in 2008 is the northward decrease in temperature of 0.2–0.3 °C in the LCL in both surveys. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

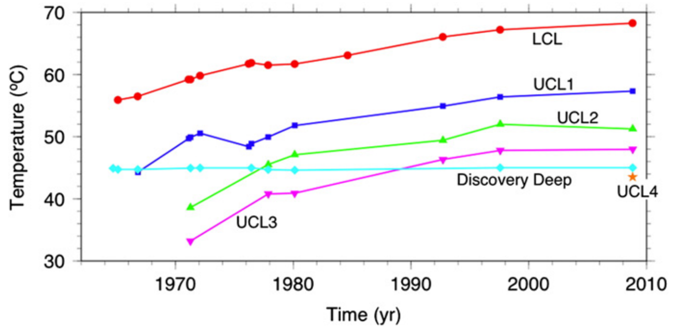

Fig. 9. The temperature of all convective layers in the Atlantis II Deep increased since 1965; the rate of increase continued to slow between 1997 and 2008. Temperature in the Discovery Deep (light blue) remained constant. We use values reported in the literature or values picked from published figures: Brewer et al. (1965); Miller et al. (1966), Ross, (1969); Brewer et al. (1969); Munns et al. (1967); Brewer et al. (1971); Bubnov et al. (1977); Schoell and Hartmann (1973); Schoell and Hartmann (1978); Hartmann (1980); Monin and Plakhin (1982); Blanc and Anschutz (1995); Anschutz and Blanc (1996); Anschutz et al. (1998); Hartmann et al. (1998a, 1999b); Schmidt et al. (2003).

1996, Table 1). To balance the heat budget, hydrothermal venting must be about 0.18 GW, which is the sum of the loss of heat to the UCL1 and the net gain in heat of the LCL. The primary uncertainty is lack of information on conductive heat exchange with the seafloor.

Our estimate of 0.18 GW appears to fall on a linear trend. Anschutz and Blanc (1996) used a different approach to the heat budget to estimate a hydrothermal flux of 0.54 GW for the period 1966–1992. Note that this estimate may be somewhat high because the authors assumed no heat flux out of the brine layers, an assumption that does not seem justified based on Fig. 9 and diffusive heat fluxes of 0.13 GW across steep temperature gradients bounding the convective layers. For 1992–1997, we computed a net heat balance for LCL of 0.15 GW from temperatures in Anschutz and Blanc (1996) and Hartmann et al. (1998b). Heat fluxes across the interface between the LCL and the UCL1 range from 0.124 GW in 1992 estimated from temperatures in Anschutz and Blanc (1996, Table 1) and an interface thickness of 3 m to 0.177–0.118 GW in 1997 using temperature differences and an interface thickness of 2–3 m from Hartmann et al. (1998a). The latter estimate assumes similar temperature offsets in the two layers caused by problems with the temperature probe in 1997 (see Hartmann et al., 1998b). The sum gives an estimated hydrothermal flux of 0.27–0.33 GW. Thus, hydrothermal flux into the LCL dropped from 0.54 in 1966–1992 to 0.27–0.33 in 1992–1997 and then to 0.18 GW in 1997–2008. Despite this decrease, the hydrothermal heat flux at Atlantis II Deep remains 3–4 times larger than that expected for a single vent in the Rainbow field on the Mid-Atlantic Ridge where German et al. (2010) estimated a flux of 0.5 GW for 10 vents.

Diffusive loss of heat through the interface between UCL4 and the LTL (0.14 GW) computed using the approach of Anschutz et al. (1999) is nearly identical to the flux of heat through the top of UCL confirming our inference that the convective heat provided by venting to the LCL propagates rapidly upwards and out of the convective layers. Heat diffusing out of UCL4 disperses into RSDW by vertical mixing through the UTL, where the density ratio is low (~2, Fig. 4) and turbulent transports become more important than diffusive in this less stratified water.

Heat is transported laterally in the convective layers as well as vertically. Fig. 7 shows that in 2008 the temperature of the upper convective layers decreased by 0.2 °C over 5–6 km whereas the temperature of the LCL remained uniform. Fig. 8 shows gradients of similar magnitude observed by previous surveys. The survey results in 2008 differ from previous results in that the temperature variation in the upper layers cannot be explained by lateral variation in heating from directly below (i.e., from the LCL). As a result, the heat budgets for these layers cannot be explained by one-dimensional heat flux models. Although our fairly simple measurements do not resolve circulation dynamics in these layers, more direct observations of water motion in this layer made by Voorhis and Dorson (1975) provide some insight.

Voorhis and Dorson (1975) investigated the nature of convection in these layers using a drifting, vertically oriented current meter suspended in UCL1. They measured up and down motions of less than a meter occurring with a period of <5 min and concluded that turbulence with a meter length scale and speeds of <1 cm/s was primarily responsible for convective mixing in the layer. They also observed larger scale vertical motions with a variety of time scales and length scales that ranged up to the layer thickness. They ascribed these to brine plumes originating at the upper and lower interfaces of UCL1. Voorhis and Dorson argued that these plumes were not numerous enough to drive convective mixing of the layer. Other recorded information indicated an even larger scale motion. They noted that the meter drifted south-eastward at about 1 cm/s and that the temperature increased by about 0.1 °C over about 3 km, a gradient that is similar in amplitude and orientation to what we observed in 2008 (Fig. 7). Voorhis and Dorson inferred that the meter was caught in a convection cell with height equal to the 30 m layer thickness and length of at least 3 km and speculated that reverse motion in the opposing limb of the cell occurred at a different level in the layer. They attribute large scale convection to unequal heating from below. The uniformity of the temperature in the LCL in 2008 (Fig. 7) indicates that the deeper layer cannot be the source of this unequal heating. Voorhis and Dorson offer another explanation. They point out that 1/20 of the volume of UCL1 lies above the flank of the basin and not above the LCL. They suggest that the lateral temperature difference between seafloor and the hot lower brine could drive overturning circulation. The persistence since at least 1971 of the horizontal temperature gradient in the upper convective layers and the lack of a horizontal gradient in the LCL at the present time supports this hypothesis.

While the temperature of four convective layers in the Atlantis II Deep has increased since 1965, the temperature of the single thick convective layer in the Discovery Deep has remained constant (Fig. 9). It is reasonable to presume that the convective layer in the Discovery Deep lost heat over this time period by diffusion across the overlying temperature gradient. Such a presumption is consistent with a 20 °C temperature drop across a 60 m thick boundary layer (Fig. 5) and with loss of heat from the Atlantis II Deep across layers with equally high temperature gradients (Fig. 9). A heat source is required to balance this loss and maintain a steady temperature in the LCL. Moreover, the double-diffusive theory for the origin of the LCL in the Discovery Deep requires a seafloor heat source at some time in the past (Turner, 1969). Following Anschutz et al. (1999), we computed a heat flux of 305 mW/m^2 through the temperature gradient in the Lower Transition Layer (Fig. 5). As suggested by Anschutz et al., this flux agrees within a factor of 2 with conductive heat flux measurements over stretched continental crust in the northern Red Sea. However, hydrothermal venting is required to explain the discrepancy between measured and theoretical conductive heat flow from young ocean crust with thin sediment cover (e.g., Anderson and Hobart, 1976, and references therein). We hypothesize an active hydrothermal vent in the Discovery Deep and suggest that this vent has remained stable long enough for convective hydrothermal heat gain to be balanced by diffusive loss through the overlying lid. The difference, then, between the two basins may not be the presence of a hydrothermal vent in the Atlantis II Deep, but a recent change in the rate of flow or the temperature of flow in the Atlantis II Deep that disturbed the balance in heat transport sometime shortly before the first observations in 1965. The slowing rate of temperature increase documented in Fig. 9 may indicate that the Atlantis II Deep is still adjusting to the change in hydrothermal venting.

4.2. Upper transition layer

As noted by Anschutz and Blanc (1996) and Anschutz et al. (1998), the most dramatic change during the last 40 years occurred at the transition between the convective brine layers and RSDW. The 2008 CTD data in Fig. 10 show that an UTL can be clearly identified at depths between about 1900 m and 1990 m by an increase of up to 5 °C and 4 psu above RSDW temperature and salinity values. The feature occurs above both basins (Fig. 5(b)). Data from the *R/V Sonne* 121 cruise in July 1997 (Winckler et al., 2001; Schmidt et al., 2003) and from the *R/V Marion Dufresne* REDSED cruise in September 1992 (Blanc and Anschutz, 1995) indicate that the layer was present with few differences from its present character. However, temperature and salinity were much lower in March and June 1976 during the cruise by the *R/V Akademik Kurchatov* (Bubnov et al., 1977; Danielsson et al., 1980; our Fig. 10) and in all data collected earlier (e.g., Munns et al., 1967; Brewer et al., 1969; Schoell and Hartmann, 1973). Whereas temperature in the deeper convective layers evolved over 42 years of observation, temperature and salinity in the 80–90 m thick UTL appear to have changed sometime between 1976 and 1992.

In contrast to temperature and salinity, oxygen in the UTL changed little between 1966 and 2008, although the data are sparse (right panels of Fig. 10). In addition to a lack of change, the oxygenation of the UTL is closer to that of overlying RSDW than deeper brine layers. Based on the 1997 data of Schmidt et al. (2003) and a few samples from our 2008 cruise, oxygen values typical of RSDW ($\sim$2 ml/l) appear to extend down in the UTL to $\sim$1960 m – within 30 m of the base of the layer (Fig. 10). Our profile of density ratio in the UTL is very close to that of RSDW outside the brine pools and increases dramatically at the base of the layer (right panels in Fig. 4). These features indicate that the vertical distribution of heat and salt in the present Upper Transition Layer probably originated by vertical mixing rather than by just molecular conduction of heat from below.

Following Anschutz et al. (1998), we propose that the present Upper Transition Layer comprises water that formed elsewhere at the seafloor and flowed into its present position over a few years at most. The vertically and laterally uniform density ratio (Fig. 4) is consistent with a new water mass that ponded above the Atlantis II and Discovery Deeps. Using the agreement in temperature between the two basins (Fig. 5), the new water mass extends at least as deep as 1970 m, the effective sill depth. Below 1970 m, mixing may have occurred between the new water mass and water comprising the steep gradient in the Lower Transition Layer. Evidence in the cooler Discovery Deep, however, suggests that molecular diffusion was more effective than mixing. Fig. 10 indicates that temperature in the layer beneath 1990 m rose between 1976 and 2008 but salinity changed little. We suggest that heat diffused – rather than mixed – downward from the new warm layer into the cooler gradient water beneath. This would explain the $\sim$1.5 °C warming of Discovery Deep transition layer water in Fig. 6 since 1976 and account for the spike in density ratio in this depth interval (Fig. 4).

The top of the new water mass coincides with the sill depth (1900 m) for the entire brine pool complex. It is likely that new water in excess of the volume between the top of the transition layers in the deep basins in 1976 and the sill at 1900 m depth spilled over into the rift to the south (Fig. 1) and may have

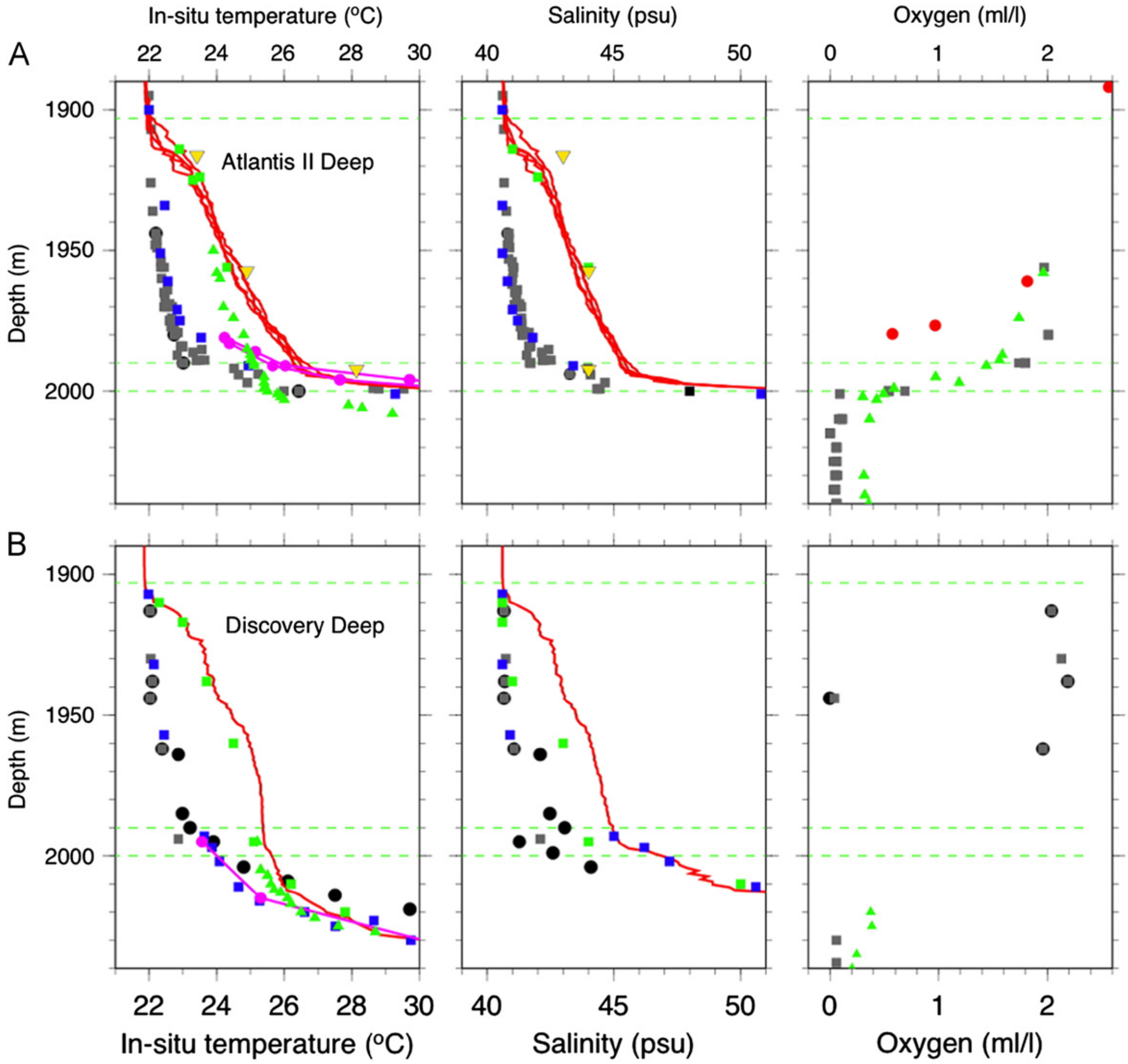

Fig. 10. Profiles through the Upper Transition Layer (1900–1990 m) of temperature (left) and salinity (middle) for the (a) Atlantis II Deep and the (b) Discovery Deep indicate that the layer appeared sometime between 1976 and 1992. Little change, however, appears to have occurred in oxygen (right) during the last 40–45 years. The pre-transition layer data include 1966 – Brewer et al. (1969: black dots), 1966 – Chain 61 data from National Ocean Data Center (gray squares), 1976 – Bubnov et al. (1977: blue squares), and 1977 – Hartmann, (1980: magenta dots and lines). Data showing an Upper Transition Layer include 1992 – Blanc and Anschutz (1995: yellow inverted triangles), 1997 – Winckler et al. (2001: green squares) and Schmidt et al. (2003: green triangles), 2008 – (SeaBird CTD in this report: red lines and oxygen laboratory analyses: red dots). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

reached the Albatross Deep. Anchutz and Blanc (1996) computed the flux across the sill due to thermal expansion of brine water.

The origin of the water in the Upper Transition Layer is unclear given that it appeared abruptly and was not derived by mixing between RSDW and brine. The possibilities include spill-over from other salt basins such as the Valdivia Deep perched on the west flank of the rift valley (Fig. 1). The temperature of brine in the Valdivia Deep increased from 1972 to 1992 consistent with hydrothermal heating (Anschutz et al., 1999). Studies by Bäcker and Schoell (1972) and Zierenberg and Shanks (1986) found that Valdivia Deep brine is lower in temperature than the Atlantis II brine, similar in salinity, and depleted in metals. We could find no measurements of oxygen in Valdivia brine water, but Schoell and Faber (1978) show that oxygen and hydrogen isotopes ratios of Valdivia brine are identical to that of RSDW. They inferred that Valdivia brine originated from flow of RSDW through salt but not through ocean crustal rock, and that the circular basin formed as a karst feature in the Miocene evaporite. Our bathymetry survey indicates a sill depth at about 1430–1450 m (Fig. 1). A 7 km long, southward-bending trough leads to the 1900 m contour of the Discovery Deep (Fig. 1; Bäcker et al., 1975).

Alternatively, the water in the Upper Transition Layer may have originated from unknown hydrothermal vents along the floor of the rift valley to the north of the Atlantis II Deep. The multibeam bathymetry map of Pautot (1983) shows a narrow fracture oriented NNW-SSE near 38°00′E, 21°32.5′N. This fracture might be an extension of the rift. A northern source is consistent with the small northward increase in temperature at 1950 m depth in the UTL that was detected in profiles from the Hobo tow-yo (top of Fig. 7). The vent might also be located along any of several scarps apparent to the east and west of the main brine deeps in Pautot's bathymetry. These scarps mark faults along which permeability might be sufficient to allow upward flow of fluid.

The sudden appearance of the Upper Transition layer emphasizes a key feature that must be considered in future studies of Red Sea brine pools: the unsteady nature of hydrothermal vents. The impetus for venting along a fault or for spillage out of Valdivia Basin is likely to be similar to that for the heat pulse that affected the temperature of the brine in the Atlantis II Deep: abrupt events in rift tectonics. The rift walls dropped to their present depths along north-south faults as the underlying continental crust cooled. This process is undoubtedly continuing at present. Movement along a fault below the salt layer may have deformed the seafloor in the Valdivia Deep enough to spill brine over its sill or to sufficiently alter the plumbing between the seafloor and Valdivia Deep to change the flux of brine. The structure beneath the rift walls is too poorly known to estimate a recurrence time for future hydrothermal events.

5. Conclusions

Brine in the Atlantis II Deep is vertically structured in paired layers comprising a vertically homogeneous convective layer (each 5–120 m thick) and a thinner interface with steep vertical temperature gradient (thickness 1–8 m) across which heat and salt are transported by diffusion. This system of convective and double diffusion layers fills both the Atlantis II and Discovery Basins up to the sill leading between the basins ($\sim$1990 m). Historical records indicate that the 80–90 m thick Upper Transition Layer above $\sim$1990 m appeared between 1976 and 1992. Limited evidence suggests that this layer is unrelated to the underlying brine pools.

The previously observed increase in temperature of convective layers in the Atlantis II Deep continues, but the rate of increase is reduced to only 0.1 °C/year. In a similar fashion, heat budgets indicate that the flux of hydrothermal venting has decreased from 0.54 GW to 0.18 GW. We suggest that the Atlantis II Deep is reaching a new equilibrium reflecting lower hydrothermal heat input; in contrast to the Discovery Deep where conductive heat flow or hydrothermal venting has remained constant. Whereas convective mixing has smoothed lateral and vertical differences observed as early as 1971 in the lower convective layer, lateral temperature gradients in the upper convective layers have not diminished in the last 40 years. Vertical and horizontal rates of heat transfer are rapid on time scales greater than 3–5 years.

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