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Heat and salt fluxes in the Atlantis II Deep (Red Sea)

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Abstract

The Atlantis II Deep is located on the Red Sea axial rift. It is a topographic depression enclosing a volume of about 17 km^3 between 1,900 m and 2,200 m depth that contains layered brines of temperatures up to 66°C and salinities up to 270‰. Previous geochemical investigations showed that the hot brines result from discharge of hydrothermal solutions that have exchanged heat and chemical components with the basaltic substratum. The last investigation of the Deep in 1992 showed that the brines occurred in four well mixed layers with the shallowest at a depth of 2,000 m. The temperature and salinity profiles describe a transition zone from 2,000 m to 1,900 m, above which Red Sea water occurs. The distribution of temperature and salt in this transition zone appears to be controlled by the topography of the Deep. The hydrographic structure of the brine column has been documented in the literature for seven time intervals between 1966 and 1992. Examination of these data shows that the system changed with time. The evolution of the entire package of brines that fill the Deep shows the following changes: (1) the temperature of the brines increased; (2) the salinity of the solutions increased; (3) the two brine layers described in 1966 still existed in 1992, but new layers appeared above them; (4) for 26 years, almost all additional heat and salt supplied in the Deep were confined to the depression and were not dispersed into the overlying seawater. The fourth point indicates that a heat and salt balance for the Deep can be calculated. The calculation of the heat that entered into the system was divided into two components: the temperature increase of the brines and the heat loss at the wall-rock; the latter component was negligible. The rate of heat input to the Deep was constant during the period considered, and amounted to 0.54×10^9 W. The salt input was also constant, and equalled 250–350 kg/s. During the period 1966–1992, heat and salt were most probably supplied by a hydrothermal solution with an average range of flow rate, temperature, and salinity of 670–1000 kg/s, 195–310°C, and 270–370‰, respectively.

Keywords: Atlantis II Deep; heat flux; brines; thermal waters

1. Introduction

The circulation of seawater through newly formed crust on mid-ocean ridges is a major feature in the heat and mass balance of the ocean. When seawater circulates through oceanic crust it is heated and its chemical composition is altered. The hot, buoyant fluid rises through fractures in the crust and exits as

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a plume, entraining fluids from the stratified ocean until it reaches the height where the buoyancy force vanishes [1]. Attempts to determine the heat flux from large scale hydrothermal areas on mid-ocean ridges have had limited success because of uncertainties in the magnitude of regional ocean currents and the local hydrography. Although numerous oceanic hydrothermal fields are now known [2], the Atlantis II Deep was the first to be discovered [3]. The Atlantis II Deep is a large, closed topographic depression filled with hot brine located at about 21°21′N on the axial zone of the young oceanic basin of the Red Sea (Fig. 1). The Deep was studied extensively during the 1970s to evaluate the economic potential of the metalliferous sediments that underlie the brine pool. Oceanic crust crops out in the area of the Deep [4], and Pliocene–Quaternary sediments and Miocene evaporites overlie the crust on the flanks of the brine pool [5]. From 1965 to 1980, many studies showed that the dense brine consists of two main layers of distinct temperature and salinity [6], called the lower and upper convective layers (i.e. LCL and UCL). The layering of the brine was interpreted as a double-diffusive convective system [7] and results in isolation of the brine pool from the overlying seawater.

The Atlantis II Deep is divided into four sub-basins which are separated by bathymetric highs that do not extend above the top of the brine pool. The area of the Deep that is covered by layered brine is limited by the 2,000 m depth contour. Previous definitions of the geometry of the Deep [6] were based on high precision mapping conducted by the Preussag company, published by Bäcker and Richter [8]. However, the size of this map is limited to the border of the Atlantis II Deep and the Discovery Deep. The more recent high resolution sea-beam map of Pautot [9] shows the closed topographic depression occupied by the Atlantis II Deep area in a greater expanse. The present study uses this map to define the area where hydrothermal fluids are topographically confined.

Data accumulated during the last 30 years allow calculation of heat and salt fluxes in the Atlantis II Deep brine system. During the 1960s and 1970s, observations of temperature and volume changes in the deepest and most concentrated brine layer indicated that high temperature hydrothermal fluids were

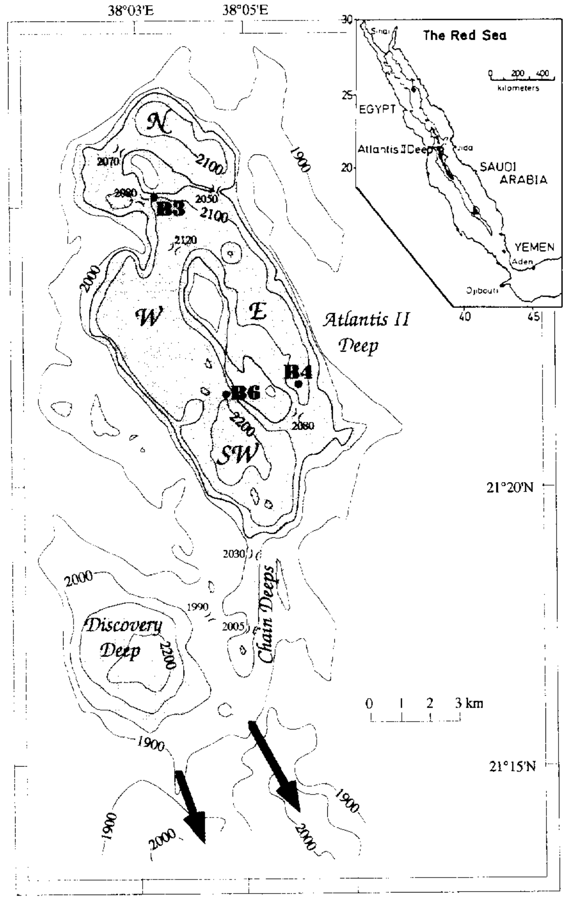

Fig. 1. Bathymetric map of the Atlantis II Deep area and location of stations B3, B4, and B6. SW = Southwest basin; W = West basin; E = East basin; N = North basin. Contour lines have been traced from seabeam investigations [9] and corrected with sound velocity calculations for Red Sea water. Corrections have also been made for the higher sound velocity in the brines. The grey area limited by the 2047 m line shows the extent of the LCL. The 2000 m contour line corresponds approximately to the top of the UCL3. The 1900 m contour line represents the maximum extent of the thermal anomaly of the Atlantis II Deep hydrothermal system in September 1992. Southwards arrows indicate the direction of the overspill of the dense brine. The maximum depth of the passes that separate the different basins of the area is indicated.

venting into the Deep [6]. Heat budget calculation of this lowest brine, and the study of fluid inclusions of the epigenetic minerals that comprise the fossil hydrothermal veins in the underlying metalliferous sediments have suggested that fluids at temperatures up to 390°C supplied the Deep [10,11]. Data collected during the REDSED cruise (September, 1992) indicated that salt added to the upper layers of the brine pool required a hydrothermal brine supply rate of about 200 l/s [12]. In the present paper we show that during the 1966–1992 period heat and salt were mostly contained within the topographic depression in which the Atlantis II Deep occurs. New inputs of heat and salt were distributed across all the brine layers, not only in the deepest one (LCL). We propose a revised heat–salt balance of the Atlantis II Deep by estimating the amount of salt and heat that was supplied to the Deep as a whole during the 1966–1992 period. The accumulation of heat and salt in the Deep allows us to determine the heat and salt fluxes of this hydrothermal field.

2. Methods

From 1965 to 1980, the Atlantis II Deep was visited regularly by oceanic research vessels but no complete hydrographic data sets were acquired during the 1980s. During the REDSED cruise (September, 1992), continuous measurements of temperature and salinity were made at three sites (Fig. 1) located in the eastern basin (station B4), the west-southwest passage (station B6), and the northern passage (station B3). Details of the methods and results have been described previously [12]. Briefly, a bathyprobe equipped with temperature, conductivity and hydrostatic pressure sensors allowed continuous recording from the sea surface to the bottom of the brine pool. Temperature recordings were calibrated with digital thermometer readings. Salinities were obtained, using a Golberg optical refractometer, from samples collected at the three stations and compared to salinity continuously measured with conductivity sensors on the bathyprobe (Fig. 2). The decrease in temperature at the bottom of station B4 allows us to demonstrate a linear relationship between conductivity and temperature at a salinity of 270‰. Likewise, the change in temperature and the negligible change in salinity in the Red Sea water column gives a similar linear relationship at 40.6‰. Then, on a conductivity versus temperature diagram, the straight lines of isosalinity for the sample for which the salinity was measured can be plotted (Fig. 2) and the intersection of the straight lines and the upper axis allowed us to define a non-linear scale for salinity.

The position of the depths separating the brine

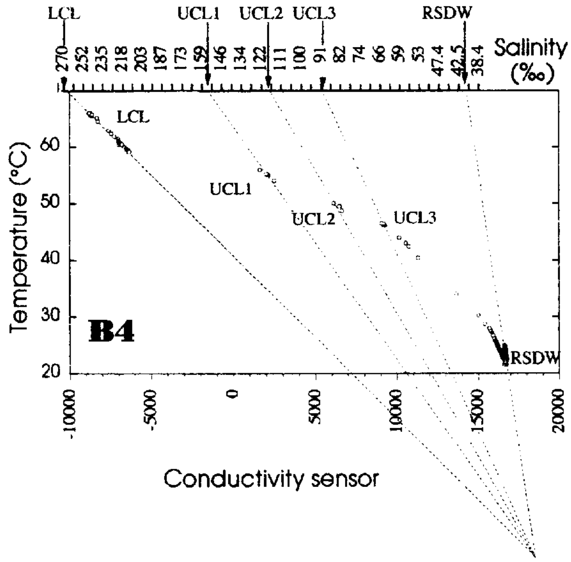

Fig. 2. Determination of the salinity of the UCL3 from a conductivity sensor data versus temperature diagram from the CTD measurements at station B4. Conductivity recordings are not calibrated because of the lack of suitable equations for the high range of temperature and salinity in the Deep, thus these values are unitless. The thermal structure of LCL at station B4 shows a straight line, which corresponds to the line of isosalinity at 270‰. The seawater column zone gives a corresponding linear relation at 40.6‰. The isosalinity lines at 153‰ and 118‰ can be drawn graphically and the intersection of the lines with the upper horizontal axis of the graph shows a non-linear scale for salinity, with $S = 2.391 \times 10^{-7}$, $CS^2 - 0.01035$ $CS + 139.77$, where CS is the conductivity sensor coordinate.

layers was calculated from the acoustic data, the bathyprobe readings, and the elevation of a pinger above the bottom. Sound velocities were determined for the seawater and the convective brine layers. The sound velocities in the brines were deduced from the extrapolation of Matthew's corrections for salinity, temperature, and pressure. For comparison, a measurement of the sound velocity in 254.9‰ brine at atmospheric pressure and 17°C gave 1762 m/s [13], compared to extrapolation of Matthew's correction to these conditions, which gave a velocity of 1753 m/s. Therefore, errors in the sound velocity in brines are probably less than 1%. The precision of the brine pool thicknesses depended on the precision of the bathyprobe position given by the pinger relative to the bottom. The maximal error was ±0.5 m. The precision of the depth of the brine interface depended mainly on the precision of the absolute depth of the bottom (i.e., on the error in the sound velocities). Between 2000 and 2200 m, the accuracy of the absolute depth was limited to ± 3 m. Here, the bottom is defined as the reflector of acoustic waves, which may be a few metres below the actual sediment–water interface because of the flocculant nature of the metalliferous sediment.

3. Results

3.1. Hydrographic data

The principal result of the September 1992 cruise was the discovery of two new brine layers overlying

Table 1
List of parameters used for the heat and salt fluxes calculations

Brine	S (mean) (‰)	T (mean) (°C)	Depth (± 3 m)	A (km^2)	h (m)	V (km^3)	Cp (mean) ($J \cdot kg^{-1} \cdot °C^{-1}$)	ρ(mean) $kg.m^{-3}$	$M \times 10^{12}$ (kg)	α(mean) (10^{-4} °C^{-1})
1992 September										
LCL	270.0	66.05			76.2	3.94	3221	1182	4.66	5.60
			2047	51.7						
UCL1	153.0	54.92			25.2	1.41	3545	1095	1.54	4.91
			2022	60.2						
UCL2	118.0	49.42			16.4	1.02	3660	1071	1.09	4.52
			2004	63.6						
UCL3	90.3	46.32			3.3	0.22	3758	1052	0.23	4.32
			2000	65.3						
TZ	45.5	23.60			97.0	11.00	3916	1032	11.35	
			1900	142.0						
RSDW	40.6	21.60					3935	1029		
1977 November										
LCL	266.5	61.50			76.2	3.94	3231	1180	4.65	5.32
			2047	51.7						
UCL1	143.8	49.93			25.2	1.41	3574	1090	1.54	4.57
			2022	60.2						
TZ	55.0	29.30			62.0	4.82	3885	1034	4.98	
			1960	101.0						
RSDW	40.6	21.60					3935	1029		
1971 April										
LCL	266.3	59.17			76.2	3.94	3233	1183	4.66	5.18
			2047	51.7						
UCL1	141.1	49.83			25.2	1.41	3583	1089	1.54	4.57
			2022	60.2						
TZ	50.0	26.00			30.0	2.00	3901	1032	2.06	
			1992	71.7						
RSDW	40.6	21.60					3935	1029		
1966 November										
LCL	262.8	56.49			76.2	3.94	3248	1183	4.66	5.01
			2047	51.7						
UCL1	132.0	44.26			25.2	1.41	3612	1087	1.53	4.16
			2022	60.2						
TZ	50	26			30.0	2.00	3901	1032	2.06	
			1992	71.7						
RSDW	40.6	21.60					3935	1029		

The surface area A of the brine interfaces and the volume V of the brine pool were calculated using the seabeam bathymetric map of Pautot [9]. The area was obtained graphically. The volume of the depression was obtained from measurements of areas at twelve depths below 2000 m and four depths between 1900 m and 2000 m. T comes from the data obtained from 1966 to 1992; the density, ρ, was calculated for the temperature and salinity of the brines using data for NaCl aqueous solutions [19]; M is the mass of the brine layers; the numerical values of c_p that correspond to the saline and thermal conditions of the brines were calculated from [20]; values of the coefficient of thermal expansion α are from [19] and correspond to pure water at the temperature of the brine.

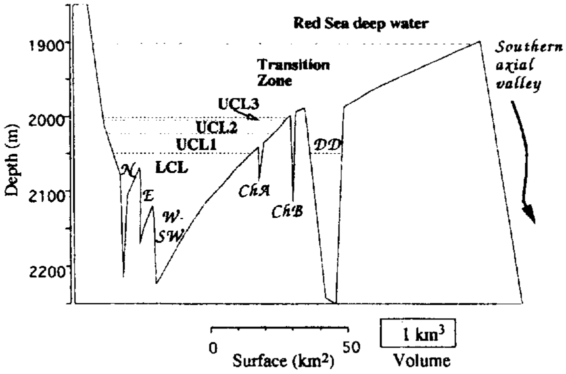

Fig. 3. Schematic vertical 2D projection of the Atlantis II Deep pool showing the different brine layers, their volume, and the surface of the interfaces. The horizontal scale shows the surface of horizontal planes at each depth. N = Northern basin; E = Eastern basin, W-SW = West and Southwest basins; ChA = Chain A Deep; ChB = Chain B Deep; DD = Discovery Deep.

the previously discovered upper brine [12]. The temperature and the conductivity showed similar patterns at the three locations. From the bottom to the top, five zones could be distinguished (Fig. 3, Table 1):

(1) The LCL was up to 150 m thick. Its upper boundary was at 2047 ± 3 m depth. The mean temperature of the upper part of this layer was 66.05°C. The temperature decreased near the brine–sediment interface at stations B4 and B6 [12]. The greatest decrease took place at site B4 in the eastern basin, where a minimum value of 59.15°C at the bottom was measured. An identical decrease in temperature was described in 1971 for station 49 [14], located close to station B4. The topography of this small depression acts as a trap for cooler, and therefore denser, fluids [14]. Chemical analyses of brine samples from these deep levels effectively showed there was no salinity gradient corresponding to the temperature gradient at B4. The salinity of the twelve samples of the LCL was the same (S = 270 ± 3‰).

(2) UCL1 is the thickest intermediate layer (25.2 ± 0.3 m), with boundaries at 2021 ± 3 and 2047 ± 3 m. The temperature was 55.20°C in the west-southwest passage and in the eastern basin, but was lower (54.31°C) in the northern passage (B3). The salinity was 153 ± 2‰.

(3) The layer between 2004 ± 3 m and 2021 ± 3 m, named UCL2, was 16.4 ± 0.5 m thick and was divided into two. The upper layer had a temperature of 49.50°C at B4 and B6, and 48.63 ± 0.30°C at B3. The bottom layer was 0.4°C warmer at all the three stations. The salinity was 118 ± 2‰.

(4) The shallowest homogenous layer, named UCL3, was 3.3 ± 0.5 m thick, between 2000 ± 3 m and 2004 ± 3 m depth. The temperature was 46.55°C at stations B4 and B6, and 45.83°C at B3. The salinity of the only sample collected in UCL3 was

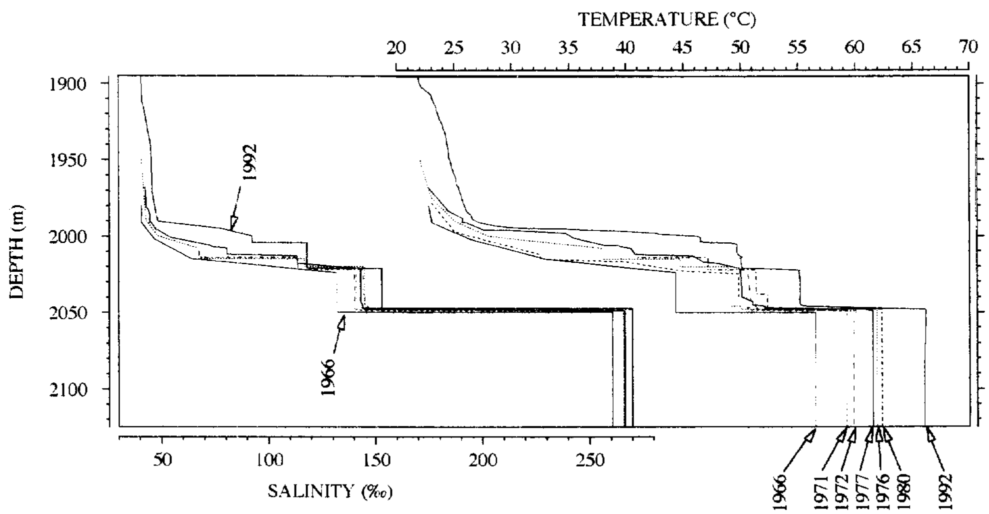

Fig. 4. Temperature and salinity vs. depth profiles from 1966 to 1992 (after [6,12,14–18]). For data given in chloride concentration, we used the equation of Danielsson et al. [42] in order to obtain salinities (S (g/l) = 1.67 Cl (g/l) + 4.02). The data for salinity given in gram/litre were converted to per mil.

107‰ (sample B6/8 in [12]). However, the salinity of UCL3, which was better constrained by the conductivity data (Fig. 2), was 90.3‰. Sample B6/8 was probably a mixture of UCL2 and UCL3 brine, which is not surprising since UCL3 is relatively thin and the sampler possibly oscillated across the UCL2–UCL3 interface because of ship and cable movements.

(5) The top of the transition zone was marked by a sudden change in temperature at 1903 ± 3 m from the temperature of the Red Sea water at this depth (21.6°C). The temperature rose to approximately 27°C at 1990 m, at which depth the gradient became very sharp, averaging 1.9°C/m. At 2000 ± 3 m, the temperature reached 45°C. A salinity gradient was observed in the transition zone with values of 40.6‰ and 90‰ at the upper and lower boundaries, respectively. A typical profile of temperature against depth is shown in Fig. 4.

The topography may explain why the shallowest temperature and conductivity anomaly was at 1900 m, because this depth corresponds to the limit where dense solutions are able to accumulate without spilling over the weir at the southward opening to the deeper and wider region of the Red Sea axial graben (Figs. 1 and 3). This may also explain why the temperature and conductivity gradients at 1900 m are higher than a few metres below. The jump in the temperature gradient at 1990 m occurred at the maximum depth where the Atlantis II Deep and Discovery Deep are connected. Above this depth, the area connected to the Atlantis II Deep increased by 15.5 km^2 (i.e., by 20%). These 15.5 km^2 were not directly above the hot brine, but above the cooler brine of the Discovery Deep [15]. Therefore, 1990 m is the depth where the heat and salt transported from the LCL is able to be dispersed laterally to a great extent.

3.2. Comparison with previous observations

The temperature, salinity, and geometry of the brines has changed with time, as documented by successive measurements in February 1965 [3], November 1966 [16], April 1971 [14], March 1972 [17], March and June 1976 [18], November 1977 [17,6], February 1980 [15], and September 1992 [12]. The temperature of the LCL increased overall by 10.13°C during 27.5 years, from 55.92°C in 1965 to 66.05°C in 1992 (Fig. 4). No reliable data exists for the UCL1 in 1965. The temperature of the UCL1 was 44.26°C in 1966 and 54.92°C in 1992 (mean value for stations B3, B4 and B6). The salinity of the LCL increased from 260‰ in 1966 to 270‰ in 1992. The salinity of UCL1 was 21‰ higher in 1992 than in 1966.

The different temperature versus depth profiles (Fig. 4) show that the LCL–UCL1 interface remained at about the same depth for 26 years and that the thickness of the UCL1 remained about 25 m. Hartmann [6] asserted an elevation of less than 2.5 m of the LCL–UCL1 interface during the 1965–1977 period, based on acoustic records. However, the echo soundings in 1965, 1971, 1972, and 1977, which he compared, were recorded at different times of the year (February, April, February, and November). Therefore, the different time of echo return from the interface may reflect a change in the sound velocity in the seawater column caused by seasonal variations in temperature and salinity of the Red Sea water, and not simply a change in the depth of the LCL–UCL1 interface. A change in the volume of the LCL in 1992 was not detected; the position of the interface (±3 m) was similar to the earlier one. One cannot exclude a weak migration of 0–5 m in the LCL–UCL1 interface position, but it was not measurable with any degree of accuracy. From 1971 to 1980 the bottom of UCL1 was about 1°C warmer than the main portion of this brine layer. This temperature step was not detected in 1992. Although UCL2 and UCL3 were not reported in previous studies, the profiles (Fig. 4) and the position of echo sounder reflections [6] show that the homogenous layers above UCL1 have probably been there since 1976. The top of the transition zone has moved upwards. It was below 1960 m depth until 1977, but at 1903 m in 1992.

4. Discussion

We have carried out a heat and salt balance calculation of the Atlantis II Deep system. We estimate the gain of heat and salt from the LCL to the top of the transition zone at 1900 m, and the heat loss by conduction through the walls of the Atlantis

II Deep. The system is considered as isovolumetric with a volume calculated from the closed basin below the 1900 m sill depth. We first quantify the heat and salt necessary to induce the T and S changes in the water column. Then we consider input of fluids.

4.1. Confinement of the system from 1966 to 1992

The top of the transition zone corresponds to the shallowest depth of temperature and salinity anomaly. The elevation of the thermal and the saline anomaly was about 60–80 m in 26 years. Some dispersion of heat and salt out of the system must have occurred just before 1992 when the anomaly reached the 1903 m level. However, since this depth remained below the depth of the weir at the southward opening from 1966 to a recent date (Fig. 4), almost all additional heat and salt supplied in the Deep remained confined to the depression.

4.2. Heat balance

Let us assume that the heat (H) that was supplied to the Deep was distributed among the different brine layers without being dispersed in the Red Sea water, as suggested by the compilation of 26 years of recordings. The heat input is the sum of heat added in all the brine layers:

$$H = \Sigma\{\rho c_p V \Delta T\} \qquad (1)$$

which originates from two components, the heat brought into the Deep by venting fluids (H^v) and the conductive heat flux through the walls and bottom of the Deep (H^c), where $H^v = H - H^c = \Sigma\{\rho c_p V\Delta T\} - H^c$; V is the volume (m^3) of the brine; ρ is the mean density (kg · m^{-3}), ΔT is the temperature change; and c_p is the specific heat (J · kg^{-1} · °K^{-1}) for each of the brine layers and the transition zone. The parameters are shown in Table 1. H^c is the expression for conductive heat flux through the walls and bottom of the Deep. H^c is positive when the heat flux is from the walls to the brine:

$$H^c = A k_b \Delta T / \Delta z \qquad (2)$$

where k_b is the thermal conductivity of the bottom rock (W · m^{-1} · °K^{-1}), $\Delta T/\Delta z$ is the temperature gradient in the rock just below the brine bottom (°K · m^{-1}), and A is the area (m^2) of the bottom in contact with brines.

Table 2
Summary of the heat and salt fluxes calculated from the hydrographic data of the 1966–1992 period

Period	Heat transfers			Salt flux
	Temperature change ($\times 10^9$ W)	Wall-rock ($\times 10^9$ W)	Venting fluids ($\times 10^9$ W)	(kg/s)
1966–1971	0.51	≤ -0.04	0.47–0.51	210
1971–1977	0.49	0.02	0.51	263
1977–1992	0.54	0.03	0.57	238

1966–1971: From 1966 to 1971, the temperature of LCL and UCL1 increased by 2.68 and 5.57°C, respectively (Table 1), whereas the temperature of the transition zone remained almost unchanged (Fig. 4). The calculated heat input is 0.51 GW (1 GW = 10^9 W) (Table 2).

The flux of heat through the bottom and walls is difficult to estimate, since the temperature gradient in the sediments underlying the brine is not known. Direct measurements of temperature of the sediments were made in 1966 [21]. At this time, the maximum temperature of the sediment column was 62°C, warmer than the LCL (56.5°C). Considering a thermal conductivity in the sediment of 0.7 W · m^{-1} · k^{-1}, a maximum conductive heat flux of 0.6–0.8 W · m^{-1} was estimated [21] (i.e. 0.03–0.04 GW) within the overall surface of the bottom of the Atlantis II Deep covered by LCL (i.e. 60 km^2). Therefore, from 1966 to 1971, the total heat input by venting brine was between 0.51 and 0.47 GW.

1971–1977: The temperature of UCL1 and LCL oscillated during this period, but an overall increase was observed. The temperature increase between 2022 and 1960 m averaged 3.3°C. The heat absorbed by the brines because of temperature increase was 0.49 GW. These calculations use mean values of heat capacity and density of the layer above UCL1.

The temperature of the LCL was similar to the temperature in the bottom sediment in 1966 [21]. Therefore, probably no heat exchange took place across the bottom of the LCL. The extension of the temperature anomaly from 1990 m to 1960 m brought about 12 km^2 of bottom area into contact with brine warmer than seawater. The bottom was covered by either muddy sediment ($k_b = 0.7$ $Wm^{-1}°K^{-1}$) or basaltic rocks ($k_b = 2$ $Wm^{-1}°K^{-1}$) [4,8]. By considering a maximum value for the temperature gradient of 1°C m^{-1}, the heat-loss by the brine was only 0.03–0.01 GW. Thus, the sum of hydrothermal heat input from 1971 to 1977 was about 0.51 GW.

1977–1992: During this time there was an overall increase in the temperature in the Atlantis II Deep system. The temperature of the brine layer between 2022 m and 1990 m depth increased by as much as 8°C, on average. The layer between 1990 m and 1900 m contains 10.3 km^3 of water which was heated by 2°C. The heat absorbed was about 0.54 GW.

The extension of the temperature anomaly to 1900 m brought more than 70 km^2 of bottom area into contact with the heated solution. Conductive heat loss must, therefore, have been important. Some heat must also have been lost advectively across the southward opening of the Deep at 1900 m, but we have no constraints on the magnitude of this heat loss. We calculated a minimum heat loss of 0.03 GW due to thermal conduction, which corresponds to a mean temperature gradient of 0.25°C · m^{-1} in a substratum of muddy sediment and basalts, and estimate that hydrothermal heat input was 0.57 GW.

The detailed temperature profiles in the Atlantis II Deep between 1966 and 1992 show that the heat that entered in the system was about 0.54×10^9 W. By allowing the volume of the LCL to increase over time, the amount of heat added when 66°C brine (maximum value of the LCL) replaced 22°C seawater is only 0.01×10^9 W for each meter increase of the LCL. This is negligible since the LCL increased by at most 5 m [6].

The estimated heat flux allows us to calculate the convective velocity in the brine. One of the parameters which characterize the fluid motion is the Rayleigh number:

$$Ra = \alpha g \Delta T h^3 / \kappa \nu \qquad (3)$$

where α is the coefficient of thermal expansion (°K^{-1}) (Table 1); g is the acceleration due to gravity (m · s^{-2}), h is the mean thickness of the layer (m); κ is the thermal diffusivity ($1.64 \pm 0.05 \times 10^{-7}$ $m^2 \cdot s^{-1}$ for Atlantis II Deep brines); and ν is kinematic viscosity ($7 \pm 1 \times 10^{-7}$ $m^2 \cdot s^{-1}$). Because of the large thickness of the brines, the Rayleigh number is of the order of 10^{16} or 10^{17}. The motion generated by the double-diffusive phenomenon is fully turbulent when the Rayleigh number is greater than about 10^7 [7]. In this case, the mean velocity v (m · s^{-1}) of brine parcels in convective motion is obtained with an approximate equation [22]:

$$v = \left(\alpha g F_H h / \rho c_p \right)^{1/3} \qquad (4)$$

where F_H is the heat flux by unit area (W · m^{-2}). The value of the heat flux that entered LCL was 0.54×10^9 W, which is about 10.5 W · m^{-2}. The heating of the LCL by 9.56°C from 1966 to 1992 indicates that 0.18×10^9 W, or 3.5 W · m^{-2}, was absorbed in this brine layer. Therefore, 7.0 W · m^{-2} entered UCL1. These heat fluxes give mean velocities of 1 cm · s^{-1} and 0.58 cm · s^{-1} for the LCL and the UCL1, respectively. A mean velocity of 0.5 cm · s^{-1} was previously measured for UCL1 in 1971 [23], which corresponds closely to our calculation.

4.3. Salt balance

The amount of salt dissolved in the Atlantis II Deep system increased between 1966 and 1992. The overall salt balance is the sum of the salt added to each brine layer:

$$S = \Sigma\{ M \Delta S \} \qquad (5)$$

where M is the mass of the brine (in kilograms) and ΔS is the salinity change (per mil). The results are as follows.

Between 1966 and 1971 the salinity of the LCL and UCL1 increased by 3.5‰ and 9.1‰, respectively, and 210 kg/s salt was added. Between 1971 and 1977 the salinity of the LCL did not change and the salinity of UCL1 increased by 2‰ at most. The greatest salt input was to the upper layers (2022–2000 m), because of the formation of homogenous layers and an increase in the salinity of the lower transition zone. The estimated salt input is about 263 kg/s.

Thanks to the precise and numerous data of chloride concentration in 1977 [6] we can define accurately the gain of salt from 1977 to 1992. Within 14.8 years, the salinities of LCL and UCL1 increased by 3.5‰ and 10‰, which corresponds to a salt gain of 67 kg/s. The development of UCL2 and UCL3 (between 2022 and 2000 m) consisted in a salt gain of 63 kg/s. In 1977, the salinity was almost at the seawater value at 1990 m and above. At 2000 m it was less than 55‰. In 1992, in contrast, the salinity of the 2000–1990 m interval was 72‰, on average. Thus, the water in this interval gained 38 kg/s salt. Between 1990 and 1900 m, the mean salinity was close to that of RSDW in 1977 (40.6‰), whereas it was 43.7‰ in 1992. The salt gain was 70 kg/s. The total salt gain was 238 kg/s for the entire Atlantis II Deep.

The salt balance for the Atlantis II Deep shows that the salt input averaged 250 kg/s. Considering the salinity increase of the LCL, 50 kg/s were gained by this brine layer, whereas 200 kg/s of salt were added to the upper layers. If we consider that the volume of LCL increased, each metre of LCL thickness increase corresponds to 20 kg/s of supplementary salt input. An increase of 5 m during the 1966–1992 period, which is a maximum value, brings the total salt input to about 350 kg/s. We conclude that the salt input was between 250 and 350 kg/s.

4.4. The source of heat and salt in the Atlantis II Deep

Our results can be summarized as follows (Table 2): The heat required to account for the temperature change of Atlantis II Deep was about 0.54 GW. At the same time the system gained at least 250 kg/s of dissolved solid. The heat that was supplied by conduction from the bottom was low in 1966 [21], in comparison with the amount of heat that entered the Atlantis II Deep system. If we assume, for the sake of argument, that all the heat was supplied by conduction after 1966, our calculations indicate that the average temperature gradient underlying the Deep would have been 13°C/m, or even higher if the heat anomaly was more local. This would have led to a positive thermal anomaly at the bottom of the LCL, which has never been measured. Therefore, conductive heat flux must be considered as negligible in comparison with advective heat flux. In fact, during the 1966–1992 period, the bottom acted mostly as a heat sink, not source.

Direct dissolution of outcropping Miocene evaporites by seawater appears to be an unlikely source of the increased salt. Rather, the high content of dissolved metals, the depletion of sulphate and magnesium [24], the oxygen isotopic composition of water, and the isotopic composition of dissolved Sr and He [25–29] suggest that the salt was provided by hydrothermal solutions that had previously interacted with the basaltic substratum. A hydrothermal source is also the best way of explaining the added heat. We conclude that a hot and salty hydrothermal brine was added to the Deep between 1966 and 1992.

The properties of the hydrothermal mineralizing fluids that supplied the pool are not known because no brine vent has been discovered. However, from our compilation of data we can estimate the average heat and salt input by this fluid. We can estimate other properties such as temperature and flow rate. The quantity of heat that is added to the Atlantis II Deep system (0.54 GW) corresponds to the heat that flows into the LCL minus the heat that flows out of the system at the 1900 m depth sill. The arriving and the escaping heat flows can be expressed as the product of the flow rate of the fluids and their heat contents. Therefore:

$$0.54 \times 10^9 = q_h H_h - q_{sw} H_{sw} \tag{6}$$

where q_h is the flow rate (in $kg \cdot s^{-1}$) of the hydrothermal brine and H_h is its heat content (in $J \cdot kg^{-1}$). H_{sw} and q_{sw} are the equivalent parameters for the escaping fluid (i.e. seawater). The equation that links q_{sw} and q_h is: $q_{sw} = \rho_{sw} \times (\Delta V + q_h/\rho_h)$, where ρ_{sw} and ρ_h represent densities and ΔV is expansion of the Atlantis II Deep brines due to the heat input (in $m^3 \cdot s^{-1}$): $\Delta V = \alpha.(0.54 \times 10^9/\rho c_p)$. Eq. (6) becomes:

$$0.54 \times 10^9 = q_h(H_h - H_{sw}\rho_{sw}/\rho_h) - \Delta V \times H_{sw} \times \rho_{sw} \tag{7}$$

For the heated brines, the average values for the coefficient of thermal expansion ($\approx 4.9 \times 10^{-4}$°C^{-1}), the density ($\approx 1098$ $kg \cdot m^{-3}$), and the heat capacity (≈ 3640 $J \cdot kg^{-1} \cdot °C^{-1}$) give a value for ΔV of 0.066 $m^3 \cdot s^{-1}$ (i.e. $\rho_{sw}\Delta V = 68$ $kg \cdot s^{-1}$). H_{sw} is about 85.9 $kJ \cdot kg^{-1}$. $\Delta V \times H_{sw} \times \rho_{sw}$ represents 5.8 $\times 10^6$ W, which is small in comparison with 0.54 $\times$ 10^9 and can therefore be neglected for the estimation of the physical properties of the incoming fluid.

Values of H_h at the in situ pressure were deduced from heat capacities of NaCl aqueous solutions computed from thermodynamic data given by Pitzer et al. [20] at 200 bar. The semi-empirical equation for the thermodynamic properties of NaCl(aq) is valid in the region 0°C ≤ T ≤ 300°C and 0 ≤ molality ≤ 6 mol · kg^{-1}. This equation may be extrapolated to higher solute molalities [20]. In a first step, the density of the hydrothermal brine ρ_h is assumed to be 1000 kg · m^{-3} and then we can calculate the temperature which corresponds to the value of H_h deduced from Eq. (7). In a second step, we apply the real value of ρ_h, corresponding to the calculated temperature and the estimated salinity and we calculate a new temperature from new c_p and H_h. The calculation converges rapidly to the values which fit in with Eq. (7). The value of the flow rate is estimated by making a number of assumptions about the salinity of the hydrothermal brine:

(1) If we assume the salinity is the same as the LCL (270‰, NaCl molality = 6.33), then an input of 250 kg · s^{-1} of salt corresponds to a flow rate of 926 kg · s^{-1}. The most appropriate values of H_h and ρ_h given by the convergence calculation are 672 kJ · kg^{-1} and 1083 kg · m^{-3}, respectively, which correspond to a 270‰ brine at 210°C with a c_p of 3202 J · °C^{-1} · kg^{-1}. If, instead, we have an input of 350 kg · s^{-1} of salt, the flow rate is 1296 kg · s^{-1} and the most appropriate values of H_h and ρ_h are 490 kJ · kg^{-1} and 1126 kg · m^{-3}, which correspond to a 270‰ brine at 155°C with a c_p of 3192 J · °C^{-1} · kg^{-1}. This kind of calculation has been made for different salinities above 270‰. The results allow us to define two curves on a T/S diagram, one for the 250 kg · s^{-1} salt input (curve A) and the other one for the 350 kg · s^{-1} salt input (curve B) (Fig. 5). The upper limit for the curves corresponds to the halite saturation or the temperature of 300°C, above which no thermodynamic data for concentrated NaCl aqueous solution are available. The halite saturation curve is shown on the T/S diagram (Fig. 5). It crosses curve B for a fluid with a salinity of 310‰ (i.e. a flow rate of 1129 kg · s^{-1}), a temperature of 183°C (c_p = 3089 J · °C^{-1} · kg^{-1}). At 300°C, curve A remains above the halite saturation curve. At this temperature, its salinity is 360‰ (i.e. a flow rate of 694 kg · s^{-1}). Finally, we report in Fig. 5 the field of the physical properties (T, S, and q_h) of the hydrothermal fluid which supplied the Atlantis II Deep from 1966 to 1992, generating a heat flux of 0.54 GW and a salt flux of 250–350 kg · s^{-1}.

(2) The flow rate can also be estimated from the

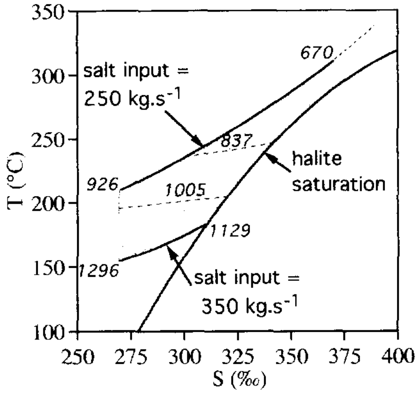

Fig. 5. Temperature vs. salinity diagram showing the field of physical properties of the hydrothermal fluid that supplied 0.54 GW and 250–350 kg/s of salt in the Atlantis II Deep from 1966 to 1992. The field is limited by the 250 kg/s and the 350 kg/s salt input curves, which have been deduced from Eq. (7). It is also limited by the halite saturation curve (after [20]) and the 270‰ salinity line. The numbers in italics represent the flow rates of the hydrothermal solution (in kg/s). The lines of equal flow rate at 1005 kg/s and 837 kg/s and the point at 670 kg/s represent the range of T/S for a water/rock ratio for heat exchange with basaltic substratum of 3, 2.5, and 2, respectively (see text).

heat content of hot basaltic rocks and the water/rock ratio between the hydrothermal fluid and the basaltic crust of the Red Sea. Using the Sr isotopic composition of the LCL brine, a water/rock ratio of 2–3 was calculated [28]. This water/rock ratio can be applied to the heat exchange. The magmatic heat associated with the formation of the oceanic crust consists of two fractions: the latent heat of crystallization of basaltic magma of about 355 × 10^3 J · kg^{-1} [30], and the heat of cooling of the rock from a temperature of about 1200°C. This latter is 1325 J · kg^{-1} · °C^{-1}, on average [31]. Therefore, 0.54 GW can be provide by cooling to 300–200°C 335 kg · s^{-1} of basaltic rock. A water/rock ratio of 2–3 for heat exchange consists then in a fluid mass (flow rate) of 670–1005 kg · s^{-1}, which are values that enter in the field defined in Fig. 5. With a maximum value of water/rock ratio of 3, the minimum temperature of the incoming fluid is 195°C (Fig. 5). The field that we have defined is consistent with the temperatures estimated from thermodynamic calculation of metal transport by the heated LCL brine [32], and with fluid ingress temperatures estimated from oxygen isotope studies of vein minerals in sediments of the Southwest basin [11].

5. Summary and conclusions

Temperature and salinity profiles from the Atlantis II Deep between 1966 and 1992 have been used to describe the supply of heat and salt to the Deep, because heat and salt remained within the system and were not dispersed into the oceanic environment. The heat flux in the Atlantis II Deep hydrothermal field was 0.54 GW. This value is lower than the heat flux from a very active hydrothermal field on the Endeavour segment of the Juan de Fuca Ridge, estimated as 3 ± 2 GW [33], 12 ± 6 GW [34], or 9.4 ± 7.4 GW [35]. This field covers 200×400 m and contains 63 venting chimneys with temperatures up to 400°C [36], as well as diffusive venting areas. The heat flow of a single black smoker was estimated to be 60×10^6 W at 21°N on the East Pacific Rise [37] and up to 50×10^6 W on the Juan de Fuca ridge [38]. The heat flux of the Atlantis II Deep is equivalent to the flux from 10 such black smokers.

The salt flux was at least 250 $kg \cdot s^{-1}$, indicating that the flow rate of the hydrothermal brine was up to 1000 $kg \cdot s^{-1}$. Such a high rate has not been reported in the literature. Converse et al. [39] found a flow rate of 150 ± 60 $kg \cdot s^{-1}$ for venting of hydrothermal fluids onto the sea floor at 21°N on the East Pacific Rise. Flow rates were not calculated for the Endeavour segment, but the high heat fluxes suggest that the discharge must be as high or higher than the rates we calculate for the Atlantis II Deep.

Previous attempts to define the flow rate and temperature of the hydrothermal fluid of the Atlantis II Deep were based on the migration of the LCL–UCL1 interface and the evolution of the temperature of the brine [6,40,41]. Hartmann [6] estimated a minimum brine discharge of about 0.28 $m^3 \cdot s^{-1}$, based on the elevation of the LCL–UCL1 interface. He noticed, however, that an additional and still greater volume was probably lost continuously from the LCL by mixing with the upper brine layers. We calculated previously that the creation of the UCL2 and UCL3 corresponds to a brine discharge of at least 0.2 $m^3 \cdot s^{-1}$ [12]. In the present study, we have shown that the brine column must be considered as a whole. Based on the increase in salinity and temperature, we calculate that 50 $kg \cdot s^{-1}$ of salt and 0.18 GW were gained by LCL. Hence, a salt flow of 250 $kg \cdot s^{-1}$ and a heat flow of 0.54 GW imply that 200 $kg \cdot s^{-1}$ of salt and 0.36 GW migrated across the LCL–UCL1 interface into the upper layers, which acted as the main sink of the newly supplied salt and heat for the 1966–1992 period.

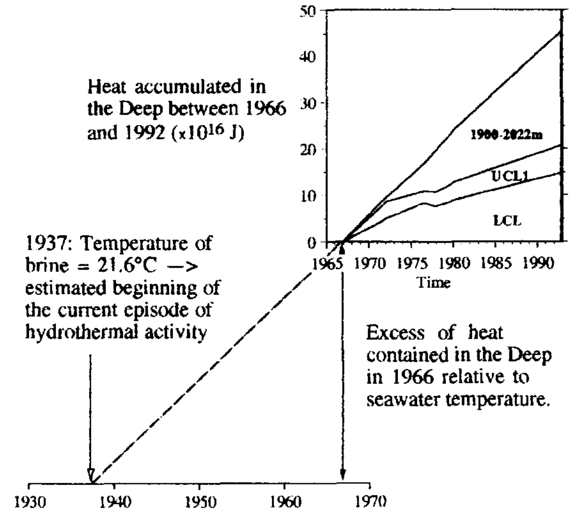

Fig. 6. Accumulated heat in the Atlantis II system from 1966 to 1992 calculated from temperature change and by taking the condition of 1966 as reference: $(T^y_x \times c^y_{px} \times \rho^y_x \times V^y) - (T^y_{1966} \times c^y_{p1966} \times \rho^y_{1966} \times V^y)$, where y is LCL, UCL1, or the 2022–1900 interval and x is 1971, 1972, 1976, 1977, 1980, and 1992. One can notice that the overall heat content increased at a constant rate (0.54 GJ/s), although individual brine layers showed fluctuations. In November 1966, the temperature of the LCL, UCL1, and transition zone corresponded to an excess of heat of about 50×10^{16} J relative to the heat content at the temperature of the Red Sea bottom water (21.6°C). A heat flux of 0.54 GW indicates that the heat has been supplied since 1937.

In 1966, the excess heat contained in the Deep relative to the heat content at seawater temperature was about 5.0×10^{17} J (Fig. 6). Assuming a constant heat flux of 0.54 GW in the recent past, the 5.0×10^{17} J have been supplied during 29 years (i.e. since 1937). Therefore, the present-day hydrothermal episode recorded from 1966 to 1992 corresponds to a period of vigorous input of heat and matter that probably begun around 1937. However, the persistence of the brine pool over thousands of years, revealed by the continuous predominance of metalliferous minerals in the Holocene sediment pile, indicates that such pulses occurred several times and, therefore, the brine has been renewed many times during the history of the Atlantis II Deep.

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