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An algorithm for estimating Absolute Salinity in the global ocean

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Abstract

To date, density and other thermodynamic properties of seawater have been calculated from Practical Salinity, S_p . It is more accurate however to use Absolute Salinity, S_A (the mass fraction of dissolved material in seawater). Absolute Salinity S_A can be expressed in terms of Practical Salinity S_p as

$$S_A = (35.16504 \text{ g kg}^{-1} / 35) S_p + \delta S_A(\varphi, \lambda, p)$$

where δS_A is the Absolute Salinity Anomaly as a function of longitude φ , latitude λ and pressure. When a seawater sample has standard composition (i.e. the ratios of the constituents of sea salt are the same as those of surface water of the North Atlantic), the Absolute Salinity Anomaly is zero. When seawater is not of standard composition, the Absolute Salinity Anomaly needs to be estimated; this anomaly is as large as 0.025 g kg^{-1} in the northernmost North Pacific. Here we provide an algorithm for estimating Absolute Salinity Anomaly for any location (φ, λ, p) in the world ocean.

To develop this algorithm we use the Absolute Salinity Anomaly that is found by comparing the density calculated from Practical Salinity to the density measured in the laboratory. These estimates of Absolute Salinity Anomaly however are limited to the number of available observations (namely 811). To expand our data set we take advantage of approximate relationships between Absolute Salinity Anomaly and silicate concentrations (which are available globally). We approximate the laboratory-determined values of δS_A of the 811 seawater samples as a series of simple functions of the silicate concentration of the seawater sample and latitude; one function for each ocean basin. We use these basin-specific correlations and a digital atlas of silicate in the world ocean to deduce the Absolute Salinity Anomaly globally and this is stored as an atlas, $\delta S_A(\varphi, \lambda, p)$. This atlas can be interpolated to the latitude, longitude and pressure of a seawater sample to estimate its Absolute Salinity Anomaly.

For the 811 samples studied, ignoring the Absolute Salinity Anomaly results in a standard error in S_A of 0.0107 g kg^{-1} . Using our algorithm for δS_A reduces the error to 0.0048 g kg^{-1} , reducing the mean square error by a factor of five. The number of sea

water samples used to develop the correlation relationship is limited, and we hope that the algorithm and error can be improved as further data becomes available.

1 Introduction

The composition of the dissolved material in seawater is not totally constant, but varies a little from one ocean basin to another. Brewer and Bradshaw (1975) and Millero (2000) point out that these spatial variations in the relative composition of seawater impact the relationship between Practical Salinity (which is essentially a measure of the conductivity of seawater at a fixed temperature and pressure) and density. The thermodynamic properties of seawater are more accurately written as functions of Absolute Salinity (as well as of temperature and pressure) rather than as functions of Practical Salinity (Millero, 1974; Millero et al., 1976b). One can make reasonable estimates of the thermodynamic properties of seawater from the concentration and known properties of the components of the solution (this simple additivity is known as Young's rule). All the physical properties of seawater as well as other multicomponent electrolyte solutions are directly related to the concentrations of the major components not the salinity determined by conductivity. Some of the variable nonelectrolytes (e.g., SiO_2 , CO_2 and dissolved organic material) do not have a conductivity signal but they do contribute to the thermodynamic properties such as density, enthalpy, entropy etc.

It is for this reason that the new thermodynamic definition of seawater (IAPWS-2008, Feistel, 2008) has the Gibbs function g of seawater expressed as a function of Absolute Salinity as $g(S_A, t, p)$ rather than as a function of Practical Salinity S_P or of Reference Salinity, S_R . Consider for example exchanging a small amount of pure water with the same mass of silicate in an otherwise isolated seawater sample at constant temperature and pressure. The conductivity is almost unchanged but the Absolute Salinity is increased and Young's rule indicates that the density, enthalpy etc. are changed in proportion to the change in Absolute Salinity. Similarly, if a small mass of say NaCl is added and the same mass of silicate is taken out of the sample, the Absolute Salin-

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ity will not have changed (and by Young's rule the density, enthalpy etc will be almost unchanged) but the Practical Salinity will have increased.

Ocean models treat their prognostic variables as possessing the "conservative" property, and the interaction of the ocean with the ice and the atmosphere already proceeds in a manner consistent with the ocean model's salinity variable being Absolute Salinity (see Jackett et al., 2006). In order to make ocean models totally consistent with TEOS-10 the models need to be initialized with Absolute Salinity and the salinity output of the models need to be compared with Absolute Salinity values derived from observations.

As a first step towards incorporating the difference between Practical Salinity S_P and Absolute Salinity S_A in oceanographic practice, Millero et al. (2008a) defined a reference composition of seawater. This reference composition defines exact mole fractions of the major components of seawater (see Table 4 of Millero et al., 2008a). Up to the accuracy of measurements to date, this reference composition is identical to that of Standard Seawater (which is surface water from a specific region of the North Atlantic). Using the most recent atomic weights, Millero et al. (2008a) calculated the Absolute Salinity of seawater of reference composition, and this salinity they called Reference-Composition Salinity, S_R . For the range of salinities where Practical Salinities are defined (that is, in the range $2 < S_P < 42$) it was shown that

$$S_R \approx u_{PS} S_P \quad \text{where} \quad u_{PS} \equiv (35.16504/35) \text{ g kg}^{-1}. \quad (1)$$

For practical purposes, this relationship can be taken as an equality since the approximation is dominated by the extent to which estimates of Practical Salinity, as determined from measurements of conductivity ratio, temperature and pressure, may vary depending on the temperature and pressure at which the measurements are made. (Although Practical Salinity was formulated so as to not depend on temperature or pressure, the algorithms used to estimate Practical Salinity at conditions other than 15°C and 0 dbar involve errors that depend on the seawater composition.) From Eq. (1), a seawater sample whose Practical Salinity S_P is precisely 35 has a Reference-Composition Salinity S_R (Reference Salinity for short) of $35.16504 \text{ g kg}^{-1}$.

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5 Millero et al. (2008a) list six reasons for introducing Reference Salinity, the last of which was to be able to use Reference Salinity as a stepping stone to Absolute Salinity, thereby being able to calculate density more accurately. Heuristically this can be thought of as reflecting the fact that some non-ionic species (such as silicate) affect the density of a seawater sample without significantly affecting its conductivity or its Practical Salinity.

The fundamental measurements required to provide a method for estimating Absolute Salinity in terms of values of Practical Salinity have been reported in Millero et al. (1976a, 1978, 2000, 2008b, 2009). The data for samples from the Southern Ocean on the CASO SR3 AA0806 voyage south of Tasmania are given in Table 2. These papers describe measurements of 811 seawater samples from around the globe at the locations shown in Fig. 1. The Practical Salinity S_P and the density ρ^{lab} of each seawater sample are measured in the laboratory at 25°C and at atmospheric pressure (assumed to be $p=0$ dbar, or an absolute pressure P of exactly 101 325 Pa) using a vibrating tube densimeter (Kremling, 1971). The Absolute Salinity of the seawater sample is estimated from the laboratory density measurement and the equation of state, essentially by solving the equation $\rho^{\text{lab}} = \rho(S_A, 25^\circ\text{C}, 0 \text{ dbar})$ for S_A . In practice the laboratory data were used to determine the density difference $\delta\rho = \rho^{\text{lab}} - \rho(S_R, 25^\circ\text{C}, 0 \text{ dbar})$ and this density difference was used with the partial derivative of density with respect to Absolute Salinity at 25°C and 0 dbar, namely $\partial\rho/\partial S_A|_{t=25^\circ\text{C}, p=0 \text{ dbar}} \approx 0.75179 \text{ kg m}^{-3}/(\text{g kg}^{-1})$, to estimate $\delta S_A = S_A - S_R$ as $\delta\rho/[0.75179 \text{ kg m}^{-3}/(\text{g kg}^{-1})]$. This is the method for estimating δS_A suggested by Millero et al. (2008a) (their Eq. 7.2).

The difference δS_A is plotted in Fig. 2a using all the data published in Millero et al. (1976a, 1978, 2000, 2008b, 2009, the Southern Ocean data of this paper). These papers have considered the correlation of various measured properties of seawater with δS_A (such as silicate, total alkalinity, total carbon dioxide and nitrate) and found that silicate correlates the best. This is fortunate as there are more measurements of silicate in the ocean data bases than either total alkalinity or total carbon dioxide.

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The reason for the good performance of silicate alone is thought to be that (a) it is itself substantially correlated with the other variables responsible for errors in using Practical Salinity to determine Absolute Salinity, (b) it accounts for a substantial fraction (of around 0.63) of the typical variations in concentrations (g kg^{-1}) of the above species and (c) being essentially non-ionic, its presence has little effect on conductivity while having a direct effect on density (Millero et al., 1976c, 2000).

The thermodynamic properties of seawater are naturally functions of Absolute Salinity rather than of Practical Salinity, and the new Thermodynamic Equation Of Seawater – 2010 (TEOS-10 for short, see McDougall et al., 2009) has algorithms for density, potential temperature, Conservative Temperature, enthalpy, entropy etc, all of which need to be called with Absolute Salinity S_A , not Practical Salinity. The algorithm of the present paper is intended to be used to estimate S_A given the “measured” values of a seawater sample’s Practical Salinity, longitude, latitude and pressure.

2 The global regressions of δS_A with silicate

15 The data in Fig. 2a, representing seawater samples from throughout the world ocean can be fitted by the simple proportional relationship with silicate (as indicated by the straight line in the figure)

$$\delta S_A/(\text{g kg}^{-1}) = (S_A - S_R)/(\text{g kg}^{-1}) = 9.824 \times 10^{-5}(\text{SiO}_2/(\mu\text{mol kg}^{-1})). \quad \text{Global} \quad (2)$$

The standard error in this fit on Fig. 1 is 0.0054 g kg^{-1} .

20 When the data in Fig. 2a are coloured by ocean basin it becomes clear that the data from different ocean basins lie either predominantly above or below the straight line fit of Eq. (2) as a function of silicate concentration. For example, the data from the North Pacific and North Indian basins clearly were on average above the straight line of Fig. 2a while the data from the Southern Ocean were clustered below the line.

25 This is not unexpected since the spatially variable relative concentrations of different constituents of seawater will not exactly co-vary with silicate.

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In order to incorporate this spatially distinct information we decided to perform different fits for the different ocean basins. Because of the dominant role of the Antarctic Circumpolar Current in transporting seawater zonally in the high southern latitudes, we posit that the zonal variation in the relative constituents of seawater may be weak and so we grouped all the data south of 30° S together and these data were fitted in a separate linear fit with silicate, as shown in Fig. 3. This fit (for latitudes south of 30° S, that is for $\lambda \leq -30^\circ$) is

$$\delta S_A / (\text{g kg}^{-1}) = 7.4884 \times 10^{-5} (\text{SiO}_2 / (\mu\text{mol kg}^{-1})). \quad \text{Southern Ocean} \quad (3)$$

The dots on Fig. 3a are the individual data points and the open circles are the values of the straight line fit Eq. (3) to the data, evaluated at the same silicate values as the data points. The error in the fit between the laboratory-determined values of $\delta S_A = S_A - S_R$ and the value from the linear fit Eq. (3) is shown in Fig. 3b. The associated standard error is 0.0026 g kg⁻¹.

The data north of 30° S in each of the Pacific, Indian and Atlantic Oceans was treated separately. In each of these three regions we constrained the fit to match Eq. (3) at 30° S and allowed the slope of the fit to vary linearly with latitude. The resulting fits were (for latitudes north of 30° S, that is for $\lambda \geq -30^\circ$)

$$\delta S_A / (\text{g kg}^{-1}) = 7.4884 \times 10^{-5} (1 + 0.3622[\lambda/30^\circ + 1]) (\text{SiO}_2 / (\mu\text{mol kg}^{-1})). \quad \text{Pacific} \quad (4)$$

$$\delta S_A / (\text{g kg}^{-1}) = 7.4884 \times 10^{-5} (1 + 0.3861[\lambda/30^\circ + 1]) (\text{SiO}_2 / (\mu\text{mol kg}^{-1})). \quad \text{Indian} \quad (5)$$

$$\delta S_A / (\text{g kg}^{-1}) = 7.4884 \times 10^{-5} (1 + 1.0028[\lambda/30^\circ + 1]) (\text{SiO}_2 / (\mu\text{mol kg}^{-1})). \quad \text{Atlantic} \quad (6)$$

These fits in the Pacific, Indian, and Atlantic Oceans north of 30° S are shown in Figs. 4–6. These fits are intended to be used from 30° S through the equator and up to the northernmost extent of these ocean basins. In the absence of density data from the Arctic Ocean, our present recommendation is that the Arctic Ocean be characterized by the same equation as the Atlantic, namely Eq. (6).

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The fitted circles in panels (a) of Figs. 4, 5 and 6 do not fall on a straight line on these plots because the fit depends on both latitude and silicate. The reason why part of a straight line is visible for the Pacific data is because much of the Pacific data is from a single latitude (see Fig. 1). It is not known why the standard deviation of the data for the Pacific and Indian Oceans are significantly larger than for the Southern Ocean. It may be that the laboratory technique for determining the conductivity and density of the samples has improved, since the Southern Ocean data was the most recent data to be measured; but this is only a conjecture at this stage. As is well-known, the silicate concentrations in the North Atlantic are quite low and it is comforting to see in Fig. 6 that the laboratory-determined values of $\delta S_A = S_A - S_R$ are also rather small there.

For each of the Southern, Pacific and Indian data sets we also performed fits that allowed an offset of δS_A at zero silicate. In no case did this significantly improve the fits. We have also plotted the residuals as functions of pressure and of latitude and detect no obvious trend in either plot.

We now have a “model” for estimating the Absolute Salinity for data from the major ocean basins. One needs to know the Practical Salinity, the location of the sample (its pressure, its latitude and which ocean basin it is from) and the silicate concentration of the sample. Having these pieces of information, one can use the appropriate equation from Eqs. (3) to (6) to calculate $\delta S_A = S_A - S_R$ for the seawater sample. This we have done for the 811 samples for which we also have the laboratory-determined values of $\delta S_A = S_A - S_R$. The error between the laboratory-determined values of $\delta S_A = S_A - S_R$ and the “model”-based values is shown as the scatter plot of errors in Fig. 2b. The standard error of these data is 0.0048 g kg⁻¹ which is a little less than that from the straight-line fit of Eq. (2) and Fig. 2a, namely 0.0054 g kg⁻¹. This improvement amounts to a reduction in error variance of 21% ($(0.0048/0.0054)^2 = 0.79$) and has been obtained by having different fits in the different ocean basins.

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3 Interpolation of the silicate atlas

We now use these correlations between $\delta S_A = S_A - S_R$ and the silicate concentration to develop a practical algorithm that can be used by oceanographers to estimate Absolute Salinity, given the Practical Salinity and the location of a seawater sample. We do this by utilizing the global atlas of Gouretski and Koltermann (2004) for (among other properties) silicate. We first use the above four Eqs. (3–6) to replace all the silicate data in the world ocean with values of Absolute Salinity Anomaly $\delta S_A = S_A - S_R$. Given the location of an observation in space, that is, the latitude λ , longitude φ , and pressure p , we interpolate the global atlas values of $\delta S_A = S_A - S_R$ to this location and then add this value of δS_A to the Reference Salinity S_R (evaluated from Practical Salinity using Eq. 1) to find the estimated Absolute Salinity S_A . We now describe in detail how the global atlas of δS_A was formed.

The Gouretski and Koltermann (2004) atlas has its silicate field (SiO_2) at a 1/2 degree by 1/2 degree horizontal resolution at 45 pressure levels ranging from the sea surface to 6131 dbar. Unfortunately this silicate field does not cover the entire global ocean, but only 99.75% of the ocean for which other hydrographic data is defined. These missing values are here filled in by averaging over the silicate values found at the four locations in the east/north/west/south directions at a distance of 1/2 degree in latitude or longitude from the point in question. In the first instance this was done along isopycnals using precise calculations of the four neutral tangent planes in all four directions. This was performed iteratively until no further missing values needed filling, when only 0.05% of the data remained without silicate values. Apart from the Caspian Sea where silicate values are not available from the Gouretski and Koltermann atlas (the Caspian Sea is excluded from the present data set; see Millero et al., 2008c for an equation of state for these waters), the locations of the remaining missing values are all coastal and so were filled in by averaging along geopotentials. This still left 0.03% of the ocean without a silicate value. All of these were against continental boundaries and were very shallow, so these missing values were set to zero, consistent with the

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surrounding near-zero silicate data at these shallow depths.

We then sub-sampled this 1/2 degree resolution ocean at 4 degrees in both latitude and longitude in the domain $[0^\circ \text{ E}, 356^\circ \text{ E}] \times [-78^\circ \text{ N}, 90^\circ \text{ N}]$. The latitude numbers have been chosen to exactly capture the northern boundary, making the computational scheme in the latitudinal direction straightforward. Since the southernmost data is located at 78.5° S , the lower boundary at 78° S captures silicate values down to 82° S . The east/west boundary condition at the Greenwich meridian is accommodated by replicating the data along the 0° E meridian at 360° E .

The final step in building the three-dimensional look-up table is to calculate from Eqs. (3–6) the Absolute Salinity Anomaly δS_A corresponding to the silicate values in the 4-degree global ocean. This is straight-forward and provides the additive adjustment to Reference Salinity that is required to complete the conversion from Practical Salinity to Absolute Salinity. The same offset can be subtracted from Absolute Salinity to yield Reference Salinity and thus obtain Practical Salinity for the inverse of the Practical to Absolute Salinity function.

To find the value of the Absolute Salinity Anomaly δS_A of a seawater sample, the global atlas of $\delta S_A(\varphi, \lambda, p)$ is interpolated as follows. The “cube” containing the data point requiring the salinity adjustment can be identified with the simplest of arithmetic operations in $x - y$ space since the longitude and latitude grids are regular. Finding the pressure index in the vertical is made using an efficient bi-section technique that can be found in, for example, Press et al. (1992). All these operations can be vectorized. When values on the upper or lower faces of the cube are missing, these values are replaced with the mean of the valid δS_A values on these same faces. When the seawater sample is deeper than the deepest non-zero $\delta S_A(\varphi, \lambda, p)$ data in the global atlas at this (φ, λ) location, the pressure of the seawater sample is artificially deemed to be the maximum pressure of non-zero δS_A values directly above the sample and the interpolation then proceeds as normal.

In Fig. 7a we have plotted a map of the silicate data (in $\mu\text{mol kg}^{-1}$) from the Gouretski and Koltermann (2004) atlas at a pressure of 2000 dbar (20 Mpa), while in Fig. 7b

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is shown the Absolute Salinity Anomaly δS_A at the same pressure. The maximum difference between the North Atlantic and North Pacific oceans at 2000 dbar is in excess of 0.025 g kg^{-1} .

4 Special treatment of ocean boundaries

5 The North Atlantic and North Pacific Oceans are closer than four degrees of latitude or longitude apart in the vicinity of the Panama Canal and if this region was not treated in a special way, the interpolation procedure described above would interpolate the silicate data of the atlas across this boundary, whereas in fact, it should be a hard boundary across which there should be no such interpolation. The contrast in silicate values is significant between the two different oceans as can be seen in the plot of silicate shown in Fig. 7a; at a pressure of 2000 dbar the difference of silicate on either side of the Panama Canal is of the order $150 \mu\text{mol kg}^{-1}$. Thus Pacific waters should be treated as no data for interpolations in the Atlantic region and viceversa. An expanded view of the Panama region is shown in Fig. 8 where we also show the simple 6-point piecewise linear function of five straight lines in longitude and latitude (in magenta) that separates the two oceans. These six points yield an efficient test to decide if a user's location is in the North Pacific or the North Atlantic. This test is only performed when the location is near this small region of the global ocean.

10 The other water mass barrier which might potentially need special treatment is the Indonesian archipelago, but in this case there is no problem since water in the Pacific and Indian oceans are well mixed above 1200 m. At great depths in this region, where the water properties do become significantly different, the two oceans are separated by more than four degrees of latitude and so the issue does not arise.

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5 The Baltic Sea

Millero and Kremling (1976) have made density measurements using the vibrating tube densimeter technique on samples from the Baltic Sea. In that paper the Absolute Salinity of water from the surface North Atlantic at $S_p=35$ was thought to be 35.171 g kg^{-1} . Using our updated estimate from Millero et al. (2008a) of $35.16504 \text{ g kg}^{-1}$, we can recast Millero and Kremling's work (1976) in the form

$$\delta S_A = S_A - S_R = 0.124(1 - S_p/35.0) \text{ g kg}^{-1}. \quad (7)$$

This equation has the reasonable property that there is no compositional correction to S_A when the conductivity of a sample is such that $S_p=35$ which indicates that any Baltic Sea water has been substantially diluted with North Atlantic seawater. Equation (7) is used in our algorithm to estimate δS_A for any seawater sample from all depths inside the Baltic Sea.

6 Summary

15 Thermodynamic properties of seawater are naturally functions of Absolute Salinity rather than of Practical Salinity, and because the new algorithms for evaluating the thermodynamic properties of seawater are functions of Absolute Salinity (McDougall et al., 2009), a method is needed to estimate this type of salinity in terms of properties that are measured at sea. Here we have described an algorithm for estimating the Absolute Salinity (g kg^{-1}) of seawater from its Practical Salinity S_p as well as the latitude, longitude and pressure of the seawater sample. The estimated standard error in the resulting value of Absolute Salinity is 0.0048 g kg^{-1} which is considerably less than the standard error involved in present oceanographic practice of effectively equating Absolute Salinity to Reference Salinity (0.0107 g kg^{-1}).

20 The algorithm exploits the correlation between the difference between Absolute and Reference Salinities and the silicate concentration. The global atlas of silicate values

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of Gouretski and Koltermann (2004) has been used together with our Eqs. (3–6) to obtain a global atlas of S_A . To estimate the Absolute Salinity of a particular seawater sample, our algorithm uses interpolations in space over the data in this atlas of Absolute Salinity Anomaly to find the value of $\delta S_A = S_A - S_R$ appropriate for the location of the seawater sample. In the Baltic Sea the approach of Millero and Kremling (1976) based on the Practical Salinity of the seawater sample has been used (Eq. 7). An example of the difference between Absolute Salinity and Reference Salinity, namely the Absolute Salinity Anomaly $\delta S_A = S_A - S_R$, is shown for a meridional vertical section through the Pacific Ocean in Fig. 9.

The algorithm described in the present paper should be regarded as a first attempt at providing a practical means of estimate Absolute Salinity. Many more measurements of density and Practical Salinity on samples collected from around the globe would probably enable the residual error to be reduced in a future algorithm.

Of the 811 samples of seawater from the world ocean that have to date been analyzed for density and hence for $\delta S_A = S_A - S_R$, the standard error of δS_A is 0.0107 g kg^{-1} . This is the standard error that is incurred if one approximates the Absolute Salinity as Reference Salinity. The maximum value of $\delta S_A = S_A - S_R$ in the open ocean occurs in the Northernmost North Pacific and is approximately 0.025 g kg^{-1} . The algorithm of this paper has been used to give estimates of Absolute Salinity for the 811 seawater samples from the world ocean. These estimates of δS_A differ from the measured density-derived values of δS_A with a standard error of 0.0048 g kg^{-1} . That is, the mean square error in evaluating Absolute Salinity by the algorithm of this paper is a factor of five less than the mean square error of equating Absolute Salinity with Reference Salinity (i.e. $(0.0107/0.0048)^2 \approx 5$). Some of the remaining error of 0.0048 g kg^{-1} is due to the error in measuring density in the laboratory (perhaps a standard error of 0.0020 g kg^{-1}) and the remaining error is due to the fact that deviations from the standard relative concentrations of the constituents of seawater are not perfectly correlated with the silicate concentration.

The computer software, in both Fortran and Matlab, which evaluates Absolute Salin-

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ity S_A given the input variables Practical Salinity S_P , longitude φ , latitude λ and sea pressure p (in dbar) is available at: www.marine.csiro.au/~jackett/TEOS-10/.

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Table 1. Nomenclature.

λ	latitude, degrees north, -90° N to $+90^\circ$ N.
φ	longitude, degrees east from the Greenwich meridian, 0° E– 360° E.
g	Specific Gibbs energy (Gibbs function), $g(S_A, t, p)$, g kg^{-1} .
S_P	Practical Salinity (salinity calculated through conductivity ratio and expressed on the Practical Salinity Scale of 1978 – Unesco, 1981, 1983)
S_A	Absolute Salinity (the mass fraction of dissolved material in seawater), g kg^{-1}
S_R	Reference-Composition Salinity, g kg^{-1} . At $S_P=35$, S_R is exactly $u_{PS}S_P$ while in the range $2 < S_P < 42$, $S_R \approx u_{PS}S_P$. “Reference-Composition Salinity” can be shortened to “Reference Salinity”.
S_{SO}	Standard Ocean Absolute Salinity, $35.16504 \text{ g kg}^{-1}$, being exactly $35 u_{PS}$, corresponding to the standard ocean Practical Salinity of 35. (see Millero et al., 2008a and Feistel, 2008)
p	sea pressure, dbar
P	absolute pressure, Pa. $P = p + P_{SO}$ where P_{SO} is the standard ocean surface pressure, $P_{SO} \equiv 101325 \text{ Pa}$.
t	Celsius temperature, ITS-90, $^\circ\text{C}$
T	absolute temperature, ITS-90, K
u_{PS}	conversion factor of Practical Salinity to Reference Salinity, exactly defined as $u_{PS} \equiv (35.16504/35) \text{ g kg}^{-1}$
δS_A	$\delta S_A = S_A - S_R$, the difference between Absolute Salinity and Reference-Composition Salinity, g kg^{-1} .
ρ	density, kg m^{-3} , $\rho^{-1} = g_\rho = \partial g / \partial p _{S_A, T}$.

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Table 2. Practical Salinity S_p and silicate SiO_2 measured in the Southern Ocean on the CASO SR3 AA0806 voyage south of Tasmania at the longitudes, latitudes and pressures shown. The densities of seawater samples collected at these locations were measured in the laboratory at $t=25^\circ\text{C}$ and $p=0$ dbar. The values of $\delta\rho$ in the table are differences between the measured densities in the laboratory and those evaluated via the equation of state using the Practical Salinity at $t=25^\circ\text{C}$ and $p=0$ dbar. The Absolute Salinity Anomaly δS_A is calculated from $\delta\rho$ as $\delta\rho/[0.75179 \text{ kg m}^{-3}/(\text{g kg}^{-1})]$.

φ ($^\circ\text{E}$)	λ ($^\circ\text{N}$)	p (dbar)	S_p	$\text{SiO}_2(\mu\text{mol kg}^{-1})$	$\delta\rho(\text{kg m}^{-3})$	$\delta S_A(\text{g kg}^{-1})$
139°55'	-65°24'	2375	34.689	102.09	0.0029	0.0038
139°55'	-65°24'	2353	34.661	102.09	0.0040	0.0053
139°55'	-65°24'	2201	34.664	102.09	0.0037	0.0049
139°55'	-65°24'	2001	34.682	104.57	0.0072	0.0095
139°55'	-65°24'	1800	34.692	107.07	0.0056	0.0075
139°55'	-65°24'	1401	34.699	111.24	0.0052	0.0070
139°55'	-65°24'	999	34.691	109.98	0.0083	0.0110
139°55'	-65°24'	600	34.634	99.50	0.0073	0.0097
139°55'	-65°24'	301	34.509	84.02	0.0038	0.0051
139°55'	-65°24'	90	34.142	58.59	0.0031	0.0042
139°55'	-65°24'	10	34.128	47.76	0.0006	0.0008
139°50'	-63°21'	3830	34.696	103.21	0.0045	0.0060
139°50'	-63°21'	3783	34.711	104.03	0.0032	0.0042
139°50'	-63°21'	3401	34.749	119.70	0.0035	0.0046
139°50'	-63°21'	3101	34.739	121.34	0.0041	0.0054
139°50'	-63°21'	2801	34.700	121.33	0.0035	0.0046
139°50'	-63°21'	2202	34.735	125.42	0.0040	0.0053
139°50'	-63°21'	1300	34.762	104.77	0.0017	0.0022
139°50'	-63°21'	598	34.775	88.25	0.0018	0.0025
139°50'	-63°21'	200	34.630	81.88	0.0044	0.0058
139°50'	-63°21'	49	34.029	30.62	0.0015	0.0019
139°50'	-63°21'	5	33.909	23.61	0.0010	0.0014
139°51'	-61°21'	4388	34.834	115.54	0.0033	0.0044
139°51'	-61°21'	4203	34.767	122.59	0.0049	0.0065
139°51'	-61°21'	3801	34.694	131.71	0.0081	0.0107
139°51'	-61°21'	3401	34.707	131.71	0.0073	0.0097
139°51'	-61°21'	3001	34.745	128.82	0.0065	0.0087
139°51'	-61°21'	2200	34.730	115.59	0.0078	0.0104
139°51'	-61°21'	1501	34.749	98.61	0.0065	0.0086

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Table 2. Continued.

φ ($^\circ\text{E}$)	λ ($^\circ\text{N}$)	p (dbar)	S_p	$\text{SiO}_2(\mu\text{mol kg}^{-1})$	$\delta\rho(\text{kg m}^{-3})$	$\delta S_A(\text{g kg}^{-1})$
139°51'	-61°21'	700	34.705	82.75	0.0037	0.0050
139°51'	-61°21'	301	34.532	73.53	0.0036	0.0047
139°51'	-61°21'	69	33.944	22.80	0.0037	0.0049
139°51'	-61°21'	5	33.801	11.94	0.0016	0.0022
139°52'	-59°51'	4529	34.685	134.63	0.0059	0.0078
139°52'	-59°51'	4199	34.733	132.62	0.0085	0.0114
139°52'	-59°51'	3799	34.697	133.48	0.0078	0.0104
139°52'	-59°51'	3399	34.731	129.73	0.0040	0.0054
139°52'	-59°51'	2600	34.841	118.88	0.0039	0.0052
139°52'	-59°51'	2198	34.742	110.53	0.0058	0.0077
139°52'	-59°51'	1201	34.812	85.47	0.0058	0.0078
139°52'	-59°51'	500	34.637	74.99	0.0042	0.0055
139°52'	-59°51'	198	34.296	53.76	0.0037	0.0049
139°52'	-59°51'	51	33.839	5.97	0.0030	0.0039
139°52'	-59°51'	5	33.798	5.93	0.0019	0.0025
139°51'	-58°21'	4032	34.749	138.84	0.0054	0.0072
139°51'	-58°21'	3602	34.703	133.09	0.0082	0.0109
139°51'	-58°21'	3199	34.729	125.70	0.0026	0.0034
139°51'	-58°21'	2900	34.744	121.18	0.0083	0.0111
139°51'	-58°21'	2602	34.752	112.65	0.0063	0.0083
139°51'	-58°21'	2300	34.906	105.98	0.0079	0.0105
139°51'	-58°21'	1701	34.817	90.37	0.0052	0.0069
139°51'	-58°21'	898	34.684	77.23	0.0037	0.0049
139°51'	-58°21'	401	34.363	55.74	0.0032	0.0042
139°51'	-58°21'	149	34.020	24.28	0.0033	0.0043
139°51'	-58°21'	30	33.820	1.87	0.0003	0.0004
139°51'	-56°56'	4179	34.782	138.70	0.0076	0.0101
139°51'	-56°56'	4134	34.708	138.69	0.0065	0.0086
139°51'	-56°56'	3801	34.717	133.85	0.0067	0.0089
139°51'	-56°56'	2999	34.728	118.52	0.0054	0.0072
139°51'	-56°56'	2500	34.811	106.42	0.0047	0.0062
139°51'	-56°56'	1400	34.747	80.60	0.0070	0.0094
139°51'	-56°56'	902	34.774	72.41	0.0066	0.0088
139°51'	-56°56'	121	33.864	11.08	0.0021	0.0028
140°44'	-55°30'	4222	34.705	138.14	0.0077	0.0102
140°44'	-55°30'	3904	34.732	134.89	0.0064	0.0085
140°44'	-55°30'	3599	34.901	132.87	0.0057	0.0075
140°44'	-55°30'	3295	34.831	120.71	0.0056	0.0074
140°44'	-55°30'	2703	34.765	119.90	0.0054	0.0071
140°44'	-55°30'	2102	34.789	106.12	0.0053	0.0071
140°44'	-55°30'	1501	34.776	89.92	0.0031	0.0041
140°44'	-55°30'	897	34.826	79.95	0.0053	0.0070
140°44'	-55°30'	597	34.639	74.32	0.0046	0.0061
140°44'	-55°30'	301	34.468	58.08	0.0018	0.0024

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Table 2. Continued.

ϕ ($^{\circ}$ E)	λ ($^{\circ}$ N)	ρ (dbar)	S_p	$SiO_2(\mu mol\ kg^{-1})$	$\delta\rho(kg\ m^{-3})$	$\delta S_A(g\ kg^{-1})$
140°44'	-55°30'	71	33.751		3.06	0.0004
141°52'	-53°35'	2658	34.749		111.31	0.0075
141°52'	-53°35'	2606	34.747		110.48	0.0070
141°52'	-53°35'	2401	34.759		103.49	0.0079
141°52'	-53°35'	2199	34.766		99.38	0.0046
141°52'	-53°35'	2000	34.765		94.45	0.0044
141°52'	-53°35'	1700	34.756		87.87	0.0062
141°52'	-53°35'	999	34.641		75.41	0.0061
141°52'	-53°35'	702	34.502		66.82	0.0022
141°52'	-53°35'	401	34.272		42.35	0.0008
141°52'	-53°35'	149	33.963		15.34	0.0023
141°52'	-53°35'	31	33.764		1.58	0.0016
143°40'	-50°10'	3297	34.748		112.10	0.0072
143°40'	-50°10'	3049	34.749		100.70	0.0055
143°40'	-50°10'	2599	34.767		89.30	0.0019
143°40'	-50°10'	2300	34.740		83.60	0.0062
143°40'	-50°10'	1603	34.520		66.43	0.0045
143°40'	-50°10'	1052	34.330		27.31	0.0000
143°40'	-50°10'	698	34.511		9.08	0.0013
143°40'	-50°10'	401	34.618		4.26	0.0015
143°40'	-50°10'	53	34.598		1.37	0.0006
146°03'	-44°43'	3267	34.858		114.05	0.0001
146°03'	-44°43'	3216	34.854		112.41	0.0061
146°03'	-44°43'	2900	34.831		105.87	0.0085
146°03'	-44°43'	2600	34.778		99.74	0.0066
146°03'	-44°43'	2301	34.766		94.83	0.0077
146°03'	-44°43'	2101	34.768		88.29	0.0061
146°03'	-44°43'	1501	34.673		74.30	0.0049
146°03'	-44°43'	699	34.589		10.05	0.0024
146°03'	-44°43'	301	34.919		3.38	0.0001
146°03'	-44°43'	79	35.275		0.95	0.0010
146°03'	-44°43'	12	35.173		1.42	0.0021
146°12'	-44°23'	2346	34.756		98.23	0.0082
146°12'	-44°23'	2307	34.763		96.51	0.0040
146°12'	-44°23'	2097	34.707		90.04	0.0053
146°12'	-44°23'	1901	34.679		85.95	0.0074
146°12'	-44°23'	1697	34.612		78.58	0.0051
146°12'	-44°23'	1500	34.550		74.85	0.0049
146°12'	-44°23'	1003	34.454		31.47	0.0033
146°12'	-44°23'	701	34.572		8.18	0.0012
146°12'	-44°23'	399	34.672		4.22	0.0006
146°12'	-44°23'	120	35.016		1.77	0.0007
146°12'	-44°23'	6	34.981		0.71	0.0021

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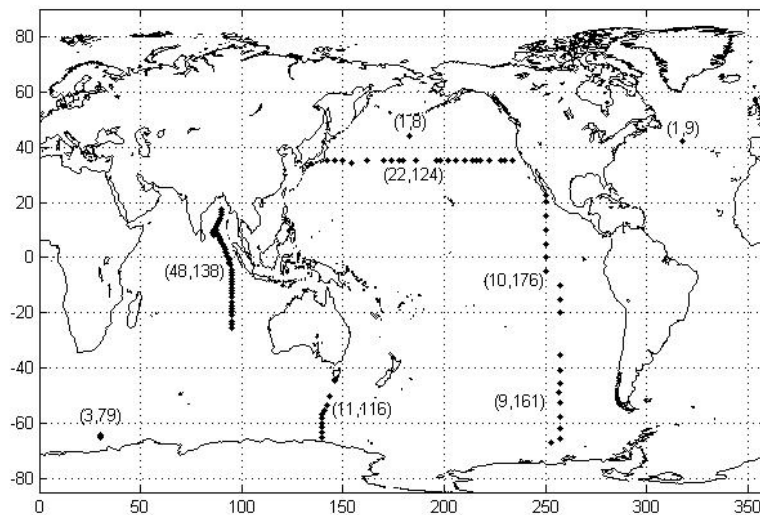


Fig. 1. Map showing the locations where the 811 seawater samples were collected whose density measurements form the basis of this paper. The first number in the brackets indicates the number of casts from which the samples were collected in each region and the second number is the number of seawater samples.

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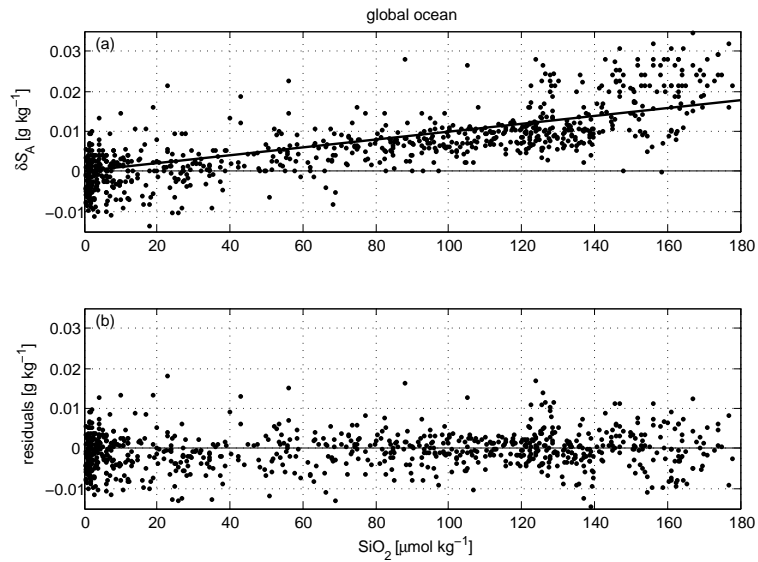


Fig. 2. (a) The laboratory-determined values of $\delta S_A = S_A - S_R$ for all 811 samples from the world ocean plotted against the silicate value of each sample. The straight-line fit to the data is given in Eq. (2) and fits these data with a standard error of 0.0054 g kg^{-1} . This straight-line fit is not the model that is adopted in this paper. The mean square of these values of δS_A is the square of 0.0107 g kg^{-1} . **(b)** The difference between the laboratory-determined value of δS_A and the model for δS_A developed in this paper represented by Eqs. (3–6). The standard error of these residuals is 0.0048 g kg^{-1} .

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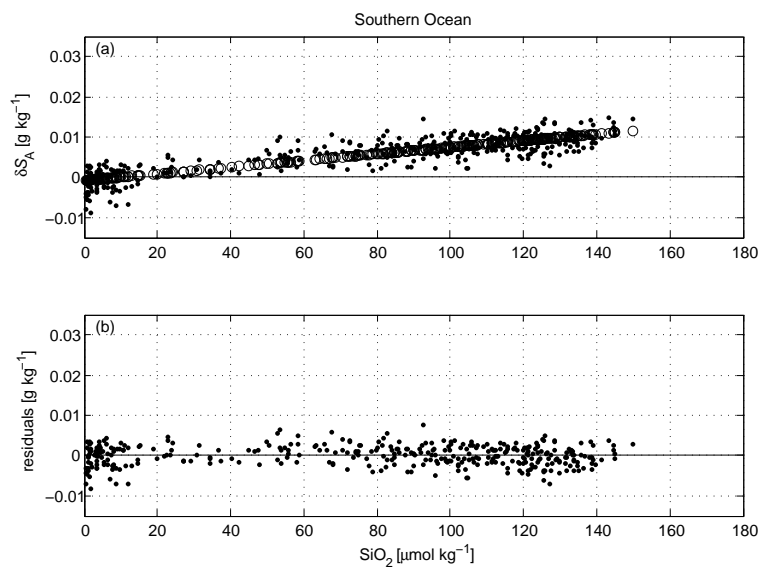


Fig. 3. (a) Laboratory-determined values of $\delta S_A = S_A - S_R$ from all longitudes and for latitudes south of 30° S plotted against the silicate value of each seawater sample. The data are the small dots and the open circles are the values obtained from the fit Eq. (3) to this data. **(b)** The residuals between the laboratory-determined values of δS_A and the values found from the fit Eq. (3).

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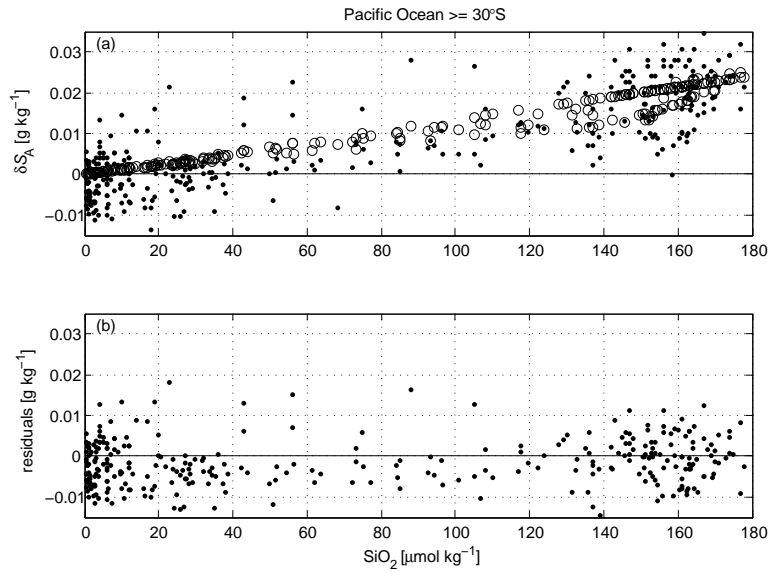


Fig. 4. (a) Laboratory-determined values of $\delta S_A = S_A - S_R$ for seawater samples from both the North and South Pacific Ocean basins north of 30° S. The data are plotted against the silicate value of each seawater sample. The data are the small dots and the open circles are the values obtained from the fit Eq. (4) to this data. **(b)** The residuals between the laboratory-determined values of δS_A and the values found from the fit Eq. (4).

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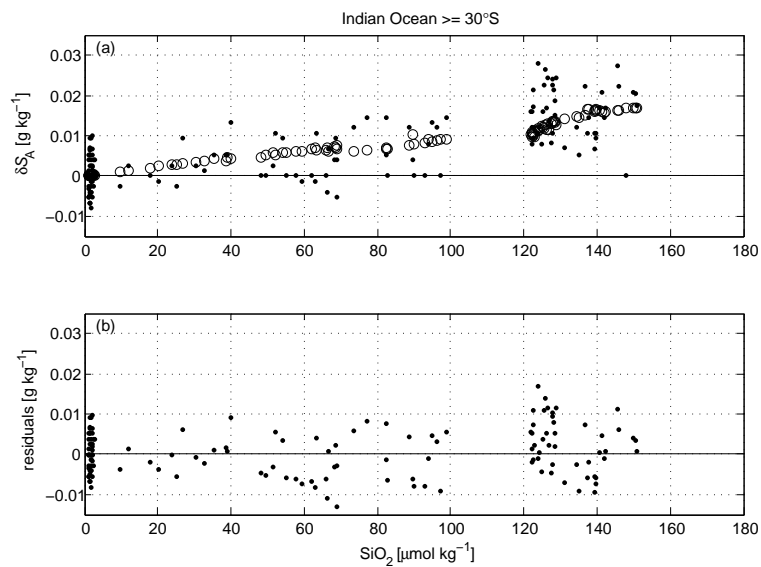


Fig. 5. (a) Laboratory-determined values of $\delta S_A = S_A - S_R$ for seawater samples from both the North and South Indian Ocean north of 30° S. The data are plotted against the silicate value of each seawater sample. The data are the small dots and the circles are the values obtained from the fit Eq. (5) to this data. **(b)** The residuals between the laboratory-determined values of δS_A and the values found from the fit Eq. (5).

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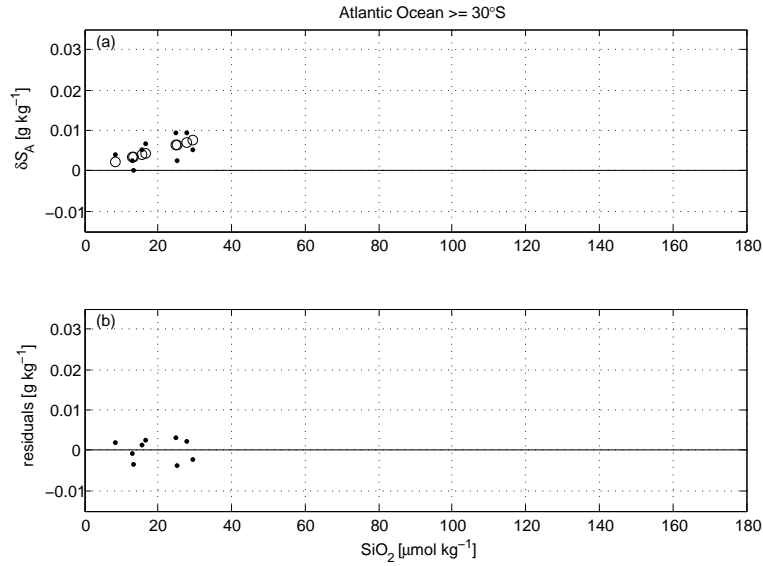


Fig. 6. (a) Laboratory-determined values of $\delta S_A = S_A - S_R$ for seawater samples from the North Atlantic Ocean. The data are plotted against the silicate value of each seawater sample. The data are the small dots and the open circles are the values obtained from the fit Eq. (6) to this data. **(b)** The residuals between the laboratory-determined values of δS_A and the values found from the fit Eq. (6).

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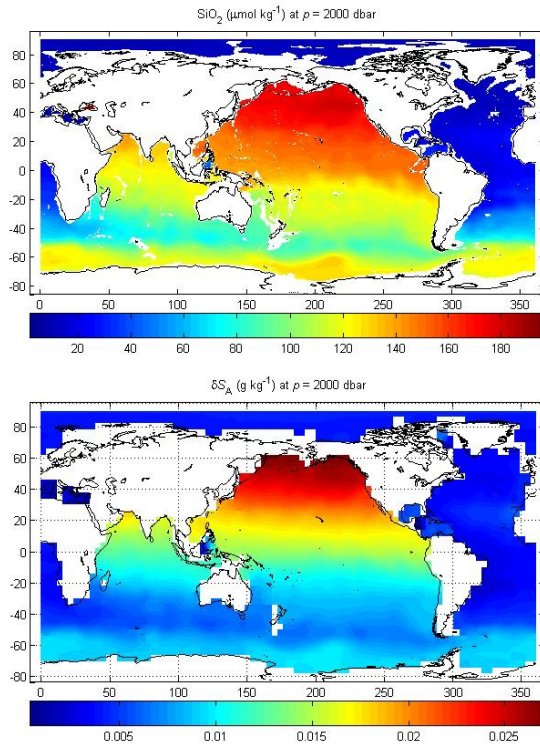


Fig. 7. (a) Silicate data (in $\mu\text{mol kg}^{-1}$) from the Gouretski and Koltermann (2004) atlas plotted at a pressure of 2000 dbar (20 Mpa). **(b)** the Absolute Salinity Anomaly δS_A at the same pressure of 2000 dbar.

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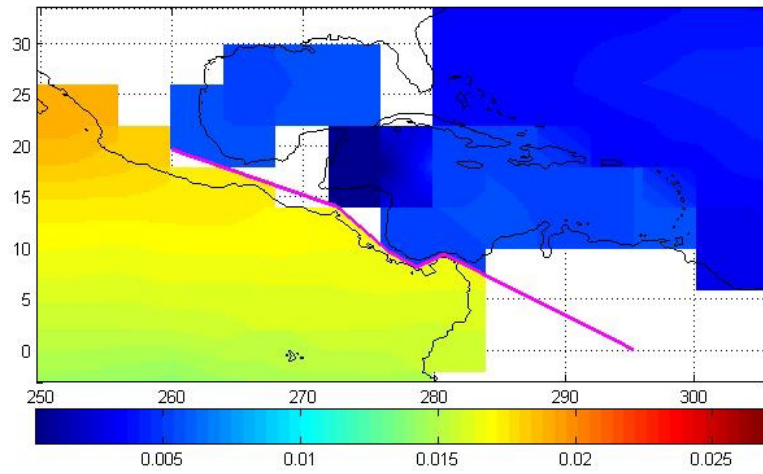


Fig. 8. An expanded view of Fig. 7b showing the series of six straight lines (in magenta) that serve to demark the North Pacific from the North Atlantic in this region.

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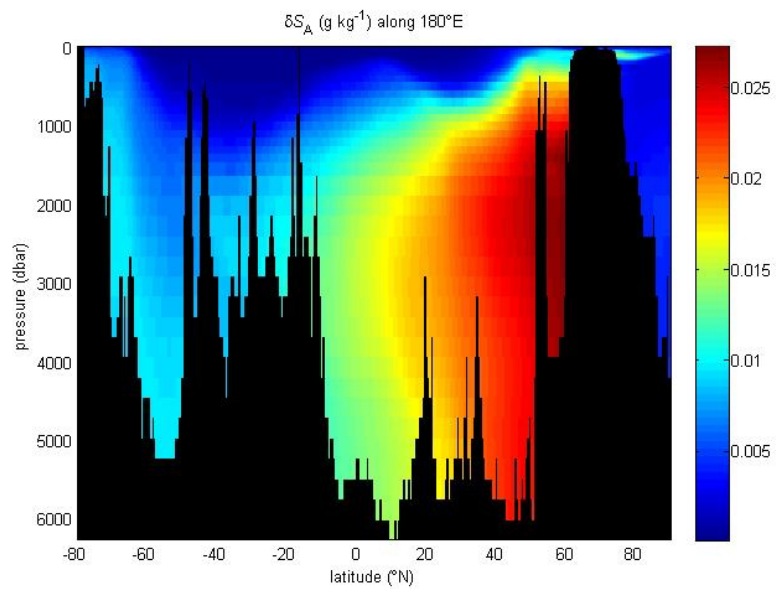


Fig. 9. A vertical section of Absolute Salinity Anomaly δS_A along 180° E in the Pacific Ocean.

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