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Kinetics of ikaite precipitation and dissolution in seawater-derived brines at sub-zero temperatures to 265 K

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Abstract

The kinetics of calcium carbonate hexahydrate (ikaite) precipitation and dissolution were investigated in seawater and seawater-derived brines at sub-zero temperatures using the constant addition experimental technique. The steady state rate of these two processes was found to be a function of the deviation of the solution from equilibrium with respect to ikaite and conformed to the same empirical rate law as the anhydrous CaCO3 polymorphs, calcite and aragonite. In addition to the saturation state of the brine with respect to ikaite, the salinity of the brine and the temperature of the reaction evidently exerted some control on the ikaite precipitation kinetics, while the dissolution kinetics of the polymorph were not noticeably influenced by these two parameters. The experimental salinity and temperature conditions were equivalent to those at thermal equilibrium between brine and ice in the sea ice cover of polar seas. Simple modelling of the CO2 system by extrapolation of the oceanic equivalent to sea ice brines showed that the physical concentration of seawater ions and the changes in ikaite solubility as a function of salinity and temperature, both inherent in the sea ice system, would be insufficient to drive the emergent brines to ikaite supersaturation and precipitation. Under these conditions, the steady state precipitation rate of ikaite was found to be fast enough for rapid formation within short time scales (days to weeks) in sea ice. The observed ikaite dissolution kinetics were also found conducive to short turn-over time scales of a few hours to a few days in corrosive solutions, such as surface seawater.

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1. INTRODUCTION

Recent field collections have documented the presence of ikaite (CaCO3·6H2O) in sea ice of the high latitude oceans (Dieckmann et al., 2008; Dieckmann et al., 2010; Rysgaard et al., 2012; Fischer et al., 2013; Rysgaard et al., 2013). The occurrence of this hydrated CaCO3 polymorph in sea ice has added an, as yet, poorly-understood abiotic component to the well-documented biological cycling of carbon by sympagic organisms (Arrigo et al., 1997; Arrigo et al., 2010; Deming, 2010).

During sea ice formation and growth, the dissolved salts in the parent seawater are excluded from the ice crystal matrix and become concentrated in the residual liquid (brine), part of which is trapped in pockets and channels in the ice matrix (Cox and Weeks, 1983; Petrich and Eicken, 2010). As a result, brine salinity increases as a function of decreasing ice temperature from the freezing point of seawater (−1.85 °C at a salinity of 35 and 1 atm total pressure) near the ice-seawater interface to colder...
temperatures in the upper parts of the ice column in the transition from summer to winter, and vice versa in the transition from winter to summer. The variation in the ionic strength and temperature of the brine inclusions during the formation–decay cycle of sea ice perturbs the thermodynamic solid–solution equilibrium of carbonate, sulphate, and chloride salts, initiating their sequential precipitation and dissolution cycles in the brine as a function of ice temperature (Marion, 2001). Ikaite is included in this sequence of authigenic solid phases in sea ice under conditions of sufficiently low concentration of dissolved carbon dioxide [here, expressed in partial pressure format (pCO2)] (Marion, 2001; Papadimitriou et al., 2013). Shifts in brine pCO2 are inherent in the sea ice formation–decay process, also as a result of the perturbation of the thermodynamic equilibria amongst the dissolved CO2 species [collectively, total dissolved inorganic carbon (C(T)) in the brine. Such shifts are inversely related to temperature and will result in a progressive brine pCO2 rise above the current atmospheric value as ice temperature drops during sea ice formation and growth (Miller et al., 2011a,b; Geiffus et al., 2012). The resulting disequilibrium relative to atmospheric pCO2 will promote transfer of dissolved CO2 to the gas phase of sea ice with concomitant lowering of the brine pCO2 (brine CO2 degassing) (Papadimitriou et al., 2004). Uptake of C(T) by the sympagic autotrophs also causes dramatic reductions in brine pCO2 to values below atmospheric equilibrium (Gleitz et al., 1995; Papadimitriou et al., 2007; Delille et al., 2007).

Sea ice and, with it, the potential areal extent of ikaite occurrence in the oceanic environment stretch over several million square kilometers in the polar oceans and sup-polar seas. As a first step towards the quantification of ikaite dynamics in sea ice, its concentration-based equilibrium solubility constant has been determined in salinity–temperature pairs matching the ice–brine thermal equilibrium down to -7.5 °C (Papadimitriou et al., 2013). The kinetics of the precipitation–dissolution cycle of this polymorph now need to be studied in similar salinity and temperature conditions before a true appreciation of its role in the polar oceanic carbon cycling can be determined. The direct field evidence for ikaite authigenesis in sea ice can hardly address the scale of ikaite production in the sea ice cover and the fate of the polymorph during sea ice decay without kinetic information. Such information will allow full integration of the ikaite cycle into existing geochemical models of sea ice, such as the FREZCHEM (Marion, 2001) and the coupled physical–biological model in Arrigo et al. (1997). Therefore, the aim of this study was to provide the first kinetic data of ikaite precipitation and dissolution at the sub-zero temperatures and high salinities typical of sea ice brines via seeded experiments at constant pCO2 (open system).

2. MATERIALS AND METHODS

2.1. The constant addition technique

The reaction kinetics of the precipitation and dissolution of ikaite were investigated in seeded open system experiments with the constant addition technique (Zhong and Mucci, 1993; Zuddas et al., 2003). The technique allows the investigation of solid–solution reaction kinetics in a steady state environment as demonstrated in Zhong and Mucci (1993). It involves the addition of a constant composition solution (reacting solution) at a constant delivery rate into a 500 mL glass reaction vessel (reactor), loaded with mineral particles (seed). The reacting solution (in this study, seawater and seawater-derived brine) was delivered in a Tygon tubing (0.64 mm ID) to the top of the reactor using a peristaltic pump (Watson Marlow 520U). The delivery rate of the reacting solution was determined gravimetrically after every experiment using distilled water. The total amount of reacting solution delivered per experiment throughout this study was 114 ± 6 g (n = 31). Synthetic ikaite was used to seed the reaction and was prepared with the method of Bishop et al. (1993) as reported in detail in Papadimitriou et al. (2013). Approximately 0.3 g of synthetic ikaite was introduced into the cold reactor, with an overall average of 0.303 ± 0.002 g (n = 31), equivalent to 2915 ± 21 μmol carbonate alkalinity and a solid to solution ratio of 2.7 g kg⁻¹ at the end of each experiment. The solution-seed mixture was continuously purged with a water-saturated CO2/N2 gas mixture with a certified CO2 composition (in ppm, equivalent to μmol/mol; uncertainty ≤5%, BOC, boc.com) through a frit at the bottom of the reactor, provided at a controlled rate (200–400 mL/min) via a CHELL C100 microprocessor controller and Hastings Mass Flow Control Valve. The certified CO2 composition of the CO2/N2 mixture was converted to partial pressure (pCO2, in μatm) using 1 atm total pressure. The reaction thus proceeded at a constant pCO2. Altering the reaction pCO2 per experiment provided a means of monitoring the extent of saturation of the reacting solution with respect to ikaite.

Seawater was used both as the reacting solution and for the preparation of the experimental brines by its variable freezing. The brines were filtered through pre-combusted (500 °C, 3 h) GF/F filters (0.7 μm, WHATMAN) and were kept stored in acid-washed glass bottles at room temperature (stock reacting solutions). The pCO2 of the prepared brines was likely to be much higher (see Section 4.3) than the atmospheric value of approximately 390 μatm due to the physical concentration of solutes present in the parent seawater during freezing. For this reason, the reacting solutions were pre-equilibrated for 10–72 h with a pCO2 close to that maintained in the reactor during the experiments but never below 390 μatm, to avoid CaCO3 precipitation in the Tygon tubing during delivery in the reactor.

During each experiment, the reacting solution, the gas hydrator, and the reactor were immersed in a constant-temperature water bath, maintained at a temperature within 0.3 °C from the freezing point of the reacting solution. This temperature was controlled by recirculation of an antifreeze cooler solution (Halfords Advanced) in a stainless steel coil at the bottom of the bath using a Cole Palmer Polystat® refrigerated chiller. To stop ice formation in the bath, the bath water was enriched with NaCl to near saturation. Exception to this pattern occurred during experiments #3A (first replicate) and #7, when the reacting solution was kept in a separate water bath at 0.8 °C and 25 °C, respectively (Table 1).
The freezing point of the reacting solution was calculated from the empirical relationship between absolute salinity (S_A, in g kg\(^{-1}\)) and temperature (t) of thermally equilibrated seawater and seawater-derived brines with ice (t ≤ −1.9 °C, S_A ≥ 34 g kg\(^{-1}\)), as well as hyposaline liquid inclusions (t > −1.9 °C, S_A < 34 g kg\(^{-1}\)) in decaying sea ice, S_A = 1000(1 − \(\frac{S}{S_A}\)) \(^{-1}\) (Assur, 1958; Leppä"ranta and Manninen, 1988; Petrich and Eicken, 2010). Absolute salinity and conductivity-based salinity (practical salinity, S) were related by S_A = 1.004715 S (Millero and Huang, 2009).

### 2.2. Steady state assessment

The aim was the determination of the steady state reaction rate, and so, it was desirable to assess the time lag for the attainment of steady state in the concentration changes in the reacting solution during the experiments. Typically, this is achieved non-intrusively by potentiometric monitoring of the change of the solution pH during solid–solution reaction with a glass electrode, to determine a plateau of constant values in the pH time series (Zhong and Mucci, 1993; Zuddas et al., 2003). This was not possible in the salinity and temperature conditions of the current experiments due to lack of suitably characterized pH buffers and the uncertainty about the electrochemical behaviour of the internal electrolyte solution of glass electrodes at sub-zero temperatures. Instead, a time series for the concentration changes in solution during solid–solution reaction was generated by varying the delivery time of a fixed amount of reacting solution (\(\frac{1}{1000} 110\) g; see Section 2.1) into the reactor, with each time point being a discrete (complete) experimental run, yielding the full suite of chemical analyses outlined in Section 2.3 below.

### 2.3. Sampling and analysis

The salinity of the reacting solutions was measured at laboratory temperature (18–26 °C) using a portable conductivity meter (WTW Cond 3110) with a WTW Tetracon 325 probe. When salinities exceeded 70, they were determined following gravimetric dilution with distilled water.

### Table 1

The composition of stock reacting solution (seawater and seawater-derived brines). All concentrations are reported in \(\mu\)mol kg\(^{-1}\) except for Ca\(^{2+}\), which is reported in mmol kg\(^{-1}\). The dry pCO\(_2\) at 1 atm total pressure used to equilibrate the stock reacting solutions prior to the kinetic experiments is in \(\mu\)atm.

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\(a\) Brine pre-equilibrated and kept at +0.8 °C during the experiment.

\(b\) Brine pre-equilibrated and kept at +24.9 °C during the experiment.

The freezing point of the reacting solution was calculated from the empirical relationship between absolute salinity (S_A, in g kg\(^{-1}\)) and temperature (t) of thermally equilibrated seawater and seawater-derived brines with ice (t ≤ −1.9 °C, S_A ≥ 34 g kg\(^{-1}\)), as well as hyposaline liquid inclusions (t > −1.9 °C, S_A < 34 g kg\(^{-1}\)) in decaying sea ice, S_A = 1000(1 − \(\frac{S}{S_A}\)) \(^{-1}\) (Assur, 1958; Leppä"ranta and Manninen, 1988; Petrich and Eicken, 2010). Absolute salinity and conductivity-based salinity (practical salinity, S) were related by S_A = 1.004715 S (Millero and Huang, 2009).
water. The water bath temperature was recorded every 15 min using data loggers (TINTYTAG AQUATIC 2 TG-4100) and also manually twice daily using type K temperature probes on a COMARK 9001 thermometer, with an overall temperature stability as ±1°C better than 0.1 °C. Samples were taken at the onset of the experimental run from the stock reacting solution and at the end, after the time lag for the attainment of steady state, from the solution in the reactor. The samples were filtered through disposable syringe filters (0.2 μm pore size, 25 mm, WHATMAN GD/X) and were used for the determination of total alkalinity (AT), CT, total dissolved calcium (Ca2+), soluble reactive phosphorus (SRP), and molybdenum-reactive silicon [hereafter, silicic acid, Si(OH)4]. The bulk synthetic material used as seed and the solids recovered from precipitation and dissolution experiments were characterized by Synchrotron X-ray Powder Diffraction on the I11 Beam Line at Diamond Light Source (Harwell Science and Innovation Campus, Oxfordshire, UK). Further sampling and analytical detail has been reported in Papadimitriou et al. (2013).

2.4. Determination of the saturation state of ikaite

The degree of saturation (Ω) of aqueous solutions with respect to a CaCO3 polymorph is given by Ω = \left[\frac{\text{Ca}^{2+}\text{[CO}_3^{2-}\right]}{K_{sp}}\right] (Millero, 1995), with quantities in brackets denoting ionic concentrations in mol kg−1 (hereafter, mol kg−1) and Ksp = ionic concentration product in solution at equilibrium with the polymorph (stoichiometric equilibrium solubility product, in mol2 kg−2). The degree of saturation is the driving force for the precipitation of the polymorph from supersaturated solutions (Ω > 1) and, conversely, for its dissolution when in contact with undersaturated solutions (Ω < 1). The determination of the Ksp of ikaite (Ksp,ikaite) in the same salinity and temperature conditions in parallel seeded closed system experiments has been reported in Papadimitriou et al. (2013). The initial saturation state of the reacting solution (Ωinitial,ikaite) is required for the modelling of the measured reaction rates. To prevent CaCO3 precipitation in the delivery tubing at low experimental pCO2 conditions (< 300 μatm), the pCO2 of the reacting solution entering the reactor was often higher than the reaction pCO2 (Tables 1 and 2). However, the direction of the reaction (precipitation or dissolution) in the reactor and its extent were driven by the reaction pCO2. Hence, the initial CO32− concentration and Ωinitial,ikaite in the reactor were estimated from the measured initial AT of the reacting solution and the pCO2 of the reaction for all experiments, coupled with the measured initial Ca2+ and the reported Ksp,ikaite (Table 3). The degree of saturation of the reacting solution with respect to ikaite at the end of each experiment (Ωfinal,ikaite) was computed from the reported Ksp,ikaite, the measured final Ca2+, and the CO32− concentration computed from the measured final AT and CT (Table 2).

All the above calculations were done by solving for pH the system of equations that describe the CO2 system of acid-base equilibrium reactions in seawater (DOE, 1994), coupling AT with CT or pCO2 as input parameters, and using the first and second dissociation constants of carbonic acid computed from the salinity and temperature functions derived from the refitting by Dickson and Millero (1987) of the measurements of Mehrbach et al. (1973). The dissociation constants of the remainder of the weak acids and bases pertinent to the oceanic CO2 system were computed from the equations in Millero (1995) and DOE (1994). The total boron concentration ([B]) required for the calculation of the contribution of borate alkalinity (AB) to AT, was calculated as a linear function of salinity (Millero, 1995) based on [B] = 0.000433 mol kg−1 at S = 35 (Lee et al., 2010). Solving the CO2 system from AT and pCO2 also required conversion of dry pCO2 to wet fCO2 (fugacity), which was done using the principles and equations in Pierrot et al. (2009). The dissolved CO2 was then calculated from the computed wet fCO2 and the CO2 solubility constant derived from the salinity and temperature function in Weiss (1974). All computations were conducted on the seawater proton scale (SWS) in mol kg−1 with the Solver routine on Microsoft Excel after extrapolation of the equations to the experimental salinity and temperature conditions.

The CO2 system in brine was also modeled using the above routine, to obtain estimates of the modification in the CO32− concentration and, from this, the change in Ωikaite in brine pockets and channels in sea ice. The model results were then combined with the kinetic results in this study, to inform on the time scale of the ikaite–solution reaction in the polar environment. The following scenarios were considered: (i) solving the CO2 system for pH with conserved surface oceanic Ca2+, AT, and CT as input parameters, equivalent to the physical concentration of solutes as a function of brine salinity and, hence, ice temperature; (ii) conserved surface oceanic AT and Ca2+ but non-conservative CT behaviour by forcing the pCO2 of the brine to the atmospheric value, equivalent to brine CO2 degassing to atmospheric equilibrium; and (iii) non-conservative Ca2+, AT and CT behaviour, by forcing the pCO2 and Ωikaite of the brine to atmospheric and ikaite–brine equilibrium (Ωikaite = 1), respectively, equivalent to CO2 degassing and ikaite precipitation from the sea ice brine. In all scenarios, the concentrations of dissolved inorganic nutrients in surface oceanic water were also conserved.

A fixed pCO2 = 390 μatm was used as an input parameter for the atmospheric value in scenarios (ii) and (iii). In scenario (iii), moreover, the system of equations was solved iteratively in two steps; in step 1, the system was solved for brine pH as in scenario (ii) for conservative surface oceanic AT and atmospheric pCO2 as input parameters; in step 2, the brine AT was modified so that the system became saturated with respect to ikaite (Ωikaite = 1) using the Ksp,ikaite reported in Papadimitriou et al. (2013) and setting the Ca2+ concentration to decline from the conservative value by an amount equivalent to half the change in AT as per stoichiometry of the ikaite precipitation reaction (see Section 3.1). For the above simulations, the brine salinity was calculated from the empirical relationship between ice temperature and brine Sx mentioned earlier in Section 2.1. Conservative concentrations were computed as linear functions of salinity. The initial input values were the concentrations measured in surface seawater (SSW) from the upper
100 m of the Weddell Sea, Antarctica, in early spring 2006, normalized to S = 35 (Papadimitriou et al., 2012): [SRP]SSW = 2.11 μmol kg\(^{-1}\), [Si(OH)\(_4\)] = 68 μmol kg\(^{-1}\), [NH\(_4\)] < 0.05 μmol kg\(^{-1}\), [CT]SSW = 2262 μmol kg\(^{-1}\), [Ar]SSW = 2369 μmol kg\(^{-1}\), with [Ca\(^{2+}\)]SSW = 0.01028 mol kg\(^{-1}\). With these input parameters and conforming to the existing empirical data set for \(K_{\	ext{sp},\text{ikaite}}\) in seawater, the model yielded kinetic studies.

2.5. Determination of the reaction rate

The net reaction (dissolution or precipitation) rate was computed from the difference between initial and final AT (AT,0 and AT,final, respectively, in μmol kg\(^{-1}\)) during the experimental time course (Δt, in h), and the stoichiometric molar ΔAT to ikaite ratio of 2 (see equilibrium reaction in Section 3.1) as (i) bulk solution rate, \(R_{\text{bulk-sln}} = \frac{[\Delta \text{AT}]_{\text{final}} - [\Delta \text{AT}]_{\text{0}}}{\Delta t}\) (in μmol kg\(^{-1}\)solution h\(^{-1}\)), (ii) rate normalized to the mass of the synthetic ikaite seed (mseed), \(R_{\text{mass}} = \frac{J[A_{\text{final}}] - [A_{\text{0}}]}{2 m_{\text{seed}}}\) (in μmol g\(^{-1}\)seed h\(^{-1}\)), and (iii) rate normalized to both mass and the estimated surface area of the synthetic ikaite seed, \(R_{\text{SA}} = \frac{J[A_{\text{final}}] - [A_{\text{0}}]}{2 \pi \cdot R_{\text{SA}} \cdot m_{\text{seed}}}\) (in μmol m\(^{-2}\)seed h\(^{-1}\)). In the above equations, J is the delivery rate of the reacting solution over the minimum of 5 days of reaction required for the achievement of steady state (0.94 ± 0.01 g h\(^{-1}\), n = 13; see Section 3.1) and SA is the geometric surface area of the synthetic ikaite seed (0.021 ± 0.010 m\(^2\) g\(^{-1}\); Papadimitriou et al., 2013). Because ikaite dissolution leads to a total alkalinity increase in solution, and vice versa during precipitation, the dissolution rates were computed and are given as positive numbers and the precipitation rates as negative.

In this study, \(R_{\text{bulk-sln}}\) was used in the simple modelling of ikaite in sea ice outlined in the previous section. The rate expressed as \(R_{\text{mass}}\) is relevant to natural settings because it is the mass rather than the surface area of ikaite that has been measured in most field studies (Dieckmann et al., 2008; Fischer et al., 2013; Rysgaard et al., 2013). Finally, \(R_{\text{SA}}\) conforms to the literature of CaCO\(_3\) kinetic studies. Hence, the ikaite kinetic rates are presented and discussed.
as $R_{SA}$ in Sections 4.1 and 4.2, while $R_{bulk-slN}$ is used in Section 4.3. In both cases, linear regression was used based on the Geometric Mean Regression theory (Ricker, 1973). Changes in the surface area of the solids during the reaction were neglected because, based on the measured $A_T$ changes, the reaction resulted in solid mass changes equivalent to less than 12% and 14% of the seed mass during dissolution and precipitation, respectively.

3. RESULTS

Bulk seed and solids recovered after the reaction experiments were identified as ikaite by X-ray powder diffraction, which indicates that ikaite was the only salt present in the solid phase during the kinetic experiments of this study. The average ($\pm$1σ) initial Ca$^{2+}$ concentration in the brines (Table 1), normalized to a salinity of 35 (10.17 ± 0.08 mmol kg$^{-1}$, $n = 7$) was comparable to the equivalent concentration in the parent seawater (10.13 ± 0.12 mmol kg$^{-1}$ at $S = 35$; Papadimitriou et al., 2013) in all but the brine used for experiment #9, in which the salinity-normalized Ca$^{2+}$ concentration was 9.91 mmol kg$^{-1}$. This indicates that in all brine batches, except that used for experiment #9, there was no measurable loss of Ca$^{2+}$ to CaCO$_3$ or CaSO$_4$ polymorphs during brine preparation by seawater freezing. The brine used for experiment #9 also had a (salinity-normalized) $A_T$ deficit relative to the remainder of the brine batches, which suggests that the Ca$^{2+}$ deficit in this brine batch was lost to CaCO$_3$ precipitation during preparation. Nonetheless, these deficits affect the carbonate ion concentration and the initial saturation state of the brine with respect to ikaite rather than the reaction mechanism. Hence, experiment #9 was considered to provide useful information on the reaction kinetics of ikaite at the maximum salinity and minimum temperature conditions of this study and was taken into account.

3.1. Time series: steady state assessment

The time required for the concentration changes resulting from the ikaite–solution reaction to reach steady state in the experimental conditions of the study was evaluated from undersaturation in seawater in experiment #1 ($S = 34$, $t = -1.1$ °C, reaction pCO$_2 = 389$ and 406 µatm) and in brine in experiment #3 ($S = 49$, $t = -2.6$ °C, reaction pCO$_2 = 389$ and 408 µatm), and from supersaturation in brine during experiment #7 ($S = 85$, $t = -4.9$ °C, reaction pCO$_2 = 22$ µatm) (Fig. 1). All concentration changes
measured during the time series experiments were commensurate with the equilibrium ikaite–solution reaction in an open system, $2\text{HCO}_3^- + \text{Ca}^{2+} + 5\text{H}_2\text{O} \rightarrow \text{CaCO}_3 \cdot 6\text{H}_2\text{O} + \text{CO}_2$, with stoichiometric molar ratios of $\Delta \text{AT} : \Delta \text{CT} = 1:1$ and $\Delta \text{AT} : \Delta \text{Ca}^{2+} = 2:1$ (Fig. 2). The time series from undersaturation (ikaite dissolution) in both seawater and brine (experiments #1 and #3; Fig. 1) indicated attainment of stable concentration (i.e., the concentration change was within the analytical and experimental variability; see next Section) after 5 days of reaction, which is commensurate with the two-week long minimum equilibration times required in closed system experiments (changing pCO$_2$) as reported in Papadimitriou et al. (2013). This was also evident in experiment #7 (ikaite precipitation), when the monitored 2 days of the precipitation reaction from supersaturation were not sufficient for the concentration changes to reach steady state (Fig. 1). Hence, experiment #7 was not used for reaction rate determination. However, the $\Delta \text{AT}$ changes observed during experiment #7 were fitted best to an empirical double exponential function of time with the boundary conditions of $[\text{AT}] = [\text{AT}]_0$ at $t = 0$ and $[\text{AT}] = [\text{AT}]_{\text{ss}}$ (steady state $\text{AT}$) at $t \rightarrow \infty$, yielding $\Delta \text{AT} = [\text{AT}]_0 - [\text{AT}]_{\text{ss}} = (1 - e^{-a_1 t}) - a_2 (e^{-b_1 t} - e^{-b_2 t})$, with $a_2$ = fitted constant. Minimizing the sum of squared residuals with the Solver routine in Excel gave $\Delta \text{AT} = -2829 (1 - e^{-0.69 t}) - 1426 (e^{-0.69 t} - e^{-1.42 t})$ ($r^2 = 0.831$, $p < 0.001$, $n = 8$) (Fig. 1a). The observed $\text{CT}$ changes during experiment #7 were also fitted to a similar empirical function of time, yielding $\Delta \text{CT} = [\text{CT}]_0 - [\text{CT}]_{\text{ss}} = -3177 (1 - e^{-0.99 t}) - 725 (e^{-0.99 t} - e^{-1.81 t})$ ($r^2 = 0.833$, $p < 0.001$, $n = 8$) (Fig. 1b). This simple empirical model indicated that, under the conditions of experiment #7, achievement of the equivalent of $\geq 98\%$ of the steady state concentration change would require $\geq 5$ days of reaction for a total reacting solution of about 110 g in the reactor. Hence, all ensuing experiments were...
conducted for 5 or 7 days, with a solution delivery rate of 0.94 ± 0.01 g hr⁻¹ (n = 13) and 0.69 ± 0.05 g hr⁻¹ (n = 18), respectively, and all reaction rates were computed from the measured ΔAₜ change for the minimum of 5 days of reaction.

3.2. Reaction kinetics of ikaite at sub-zero temperatures

A total of 31 seeded experiments were conducted for the investigation of the reaction rate of ikaite precipitation and dissolution at sub-zero temperatures in high ionic strength natural solutions (Tables 1 and 2). Detailed information on the estimated ŷ̂ₗ_initial and the measured reaction rates is given in Table 3. All concentration changes were commensurate with the equilibrium CaCO₃-solution interaction in an open system as described in Section 3.1 (Fig. 2). The replication of several experiments, i.e., #1, #2A, #2B, #3B, #5A, and #5F (n = 2), as well as #3A (n = 5), yielded an estimate of the combined analytical and experimental variability of the final Aₜ concentration (Table 2), on which the reaction rate is based. Expressed as range of the final mass per experiment, the combined analytical and experimental variability ranged from 32 to 287 mmol kg⁻¹, equivalent to 0.6–6% of the average final Aₜ at [ΔAₜ] ≈ 3500–5500 mmol kg⁻¹. The attendant uncertainty in the reaction rate (Table 3) was better than 26%, except for the very low dissolution rates in experiment #3B, where it was 81%. Overall, the direction of the reaction was effectively controlled by the pCO₂ in the reactor. The estimated ŷ̂ₗ_initial ranged from 1.02 to 3.69 in conditions of supersaturation with respect to ikaite (n = 15) and from 0.18 to 0.92 in conditions of undersaturation (n = 16) (Table 3). An uncertainty in ŷ̂ₗ_initial of up to 0.03 Ω-units was due to the analytical uncertainty in the initial Ca²⁺ concentration (0.2–2.0%; Papadimitriou et al., 2013). Additional uncertainty in ŷ̂ₗ_initial was incurred by the indeterminable computational uncertainty in the initial CO₂⁻ concentration from the extrapolation of the salinity and temperature functions that describe the oceanic CO₂ system to the sub-zero temperatures and high salinities of the experiments. The overall uncertainty is particularly evident in reaction conditions very close to equilibrium from the direction of supersaturation in experiments #3B (ŷ̂ₗ_initial = 1.03 and 1.06; n = 2) and #5C (ŷ̂ₗ_initial = 1.02). In those experiments, the calculated ŷ̂ₗ_initial indicated slight supersaturation, while the observed final ΔAₜ indicated ikaite dissolution (positive rate; Table 3), suggesting slight undersaturation. Thus, supersaturation conditions within less than 0.1 units from Ω = 1 cannot be resolved with the current combined analytical, experimental, and computational uncertainty, and will not be taken into further consideration.

4. DISCUSSION

4.1. The empirical rate law of ikaite precipitation and dissolution

The kinetics of the precipitation and dissolution of the anhydrous CaCO₃ polymorphs calcite and aragonite, determined as bulk rate changes in solution composition, have been typically related to the saturation state of the solution with respect to these polymorphs as ŷ̂ₗ_initial = k_initial(Ω – 1)ⁿ⁺ and ŷ̂ₗ_diss = k_diss(1 – Ω)ⁿ⁺, respectively (Zhong and Mucci, 1993; Morse et al., 2007). In these empirical rate laws, k = reaction rate constant (in μmol m⁻² h⁻¹) and n = reaction order, while the functions are linearized by logarithmic transformation. Although not suitable for assessment of solid–solution reaction mechanisms at the microscopic level, the empirical rate laws have been useful in environmental applications to both dilute and highly concentrated electrolyte solutions, such as seawater and brines. The current observations of the surface-area-normalized ikaite precipitation rate (Rₚₚₛ) yielded n = 1.23 ± 0.42 and log kₚₚₛ = 2.07 ± 0.23 within 1.1 ≤ ŷ̂ₗ_initial ≤ 3.7 (r² = 0.742, p < 0.001, n = 12; Fig. 3a). The equivalent analysis for the ikaite dissolution rate yielded n = 0.98 ± 0.18 and log k_diss = 2.41 ± 0.09 within 0.2 ≤ ŷ̂ₗ_initial ≤ 0.9 (r² = 0.888, p < 0.001, n = 16; Fig. 3b).

The current measurements are the first set of observations for the ikaite reaction kinetics in natural waters. To place them in the broader context of metal carbonate reaction kinetics in the aquatic environment, the current observations can be compared tentatively to the more widely available equivalent kinetic rate observations for calcite and aragonite. The literature data have been derived either in synthetic seawater or in compositionally simpler NaCl–NaHCO₃–CaCl₂ solutions of ionic strength ranging from the dilute (<0.1 m) to the highly concentrated (ionic strength ≥ 0.7 m). Given the ionic complexity of the current media, our primary comparison will be with kinetic data obtained from the reaction of anhydrous CaCO₃ polymorphs with synthetic seawater. This comparison shows that the reaction order of ikaite precipitation (n = 1.23 ± 0.42) at sub-zero temperatures is comparable to – but closer to first order reaction kinetics than – that of calcite precipitation at 5 °C (n = 1.55 ± 0.17) and 25 °C (n = 1.84 ± 0.20 and 2.22 ± 0.05) (Zhong and Mucci, 1993; Lopez et al., 2009) and aragonite precipitation at 25 °C (n = 1.90 ± 0.10; Burton and Walter, 1987). Moreover, the ikaite precipitation rate in the current hypersaline and sub-zero temperature conditions close to saturation (i.e., ŷ̂ₗ_initial ≈ 3.7) was higher than the reported calcite precipitation kinetics at above-zero temperatures in the Ω range of this study, being as high as aragonite precipitation kinetics in seawater at 25 °C. The comparison extends to the saturation conditions, although the precipitation kinetics of the anhydrous CaCO₃ polymorphs was also measured at much higher supersaturation than that of ikaite in the present study (Fig. 3a). The above comparison becomes more evident in terms of the reaction rate constant for ikaite precipitation (log kₚₚₛ = 2.07 ± 0.23), which was similar to that for aragonite in seawater at 25 °C (log kₚₚₛ = 1.61 ± 0.01; Burton and Walter, 1987) but higher, by an order of magnitude, than that reported for calcite in much warmer seawater, i.e., log kₚₚₛ = 0.20 ± 0.04 at 5 °C, and 0.33 ± 0.05 and 0.21 ± 0.13 at 25 °C (Zhong and Mucci, 1993; Lopez et al., 2009). The precipitation rate of calcite has been reported to decline by a similar amount in the presence of ionic inhibitors in the composition of seawater (e.g., magnesium and sulphate ions, orthophosphate).
In contrast, the observed dissolution kinetics of ikaite differed substantially from that of synthetic and biogenic calcite, low Mg calcite, and aragonite (Fig. 3b). The current observations yielded first order dissolution kinetics for ikaite ($n = 0.98 \pm 0.18$), which is distinctly different from the second to fourth order reported for the dissolution reaction of the anhydrous CaCO$_3$ polymorphs in seawater at 20°C (Gehlen et al., 2005). Furthermore, the reaction rate constant for ikaite dissolution ($\log k_{\text{diss}} = 2.41 \pm 0.09$) was larger than that for sedimentary biogenic calcite and aragonite explored for their dissolution kinetics in warm synthetic seawater in Gehlen et al. (2005). Furthermore, the reaction rate constant for ikaite dissolution ($\log k_{\text{diss}} = 2.41 \pm 0.09$) was larger than that for sedimentary biogenic calcite and aragonite explored for their dissolution kinetics in warm synthetic seawater in Gehlen et al. (2005) (log $k_{\text{diss}} = 0.57 - 1.30$, after unit conversion). It appears, therefore, that the ikaite dissolution rate in cold seawater and seawater-derived brines is much higher than that of calcite and aragonite in warm seawater.

4.2. The effect of ionic strength and temperature

Temperature variation affects solid–solution reaction kinetics via the temperature effect on the reaction rate constant (Morse et al., 2007). In addition, both the stoichiometric solubility of ikaite and the equilibrium speciation of dissolved CO$_2$ will be affected by the temperature of the solid–solution reaction and the ionic strength of the medium (Mucci, 1983; DOE, 1994; Millero, 1995; Papadimitriou et al., 2013). The temperature–salinity–pCO$_2$ conditions of the current experiments were chosen to reflect the temperature–salinity co-variation of thermally equilibrated sea ice brines (Assur, 1958) and the lower part of the pCO$_2$ range reported in sea ice to values well below atmospheric equilibrium (Gleitz et al., 1995; Delille et al., 2007; Papadimitriou et al., 2007). Such low pCO$_2$ is, evidently, most favourable to ikaite precipitation (Papadimitriou et al., 2013) and, consequently, to its investigation.

The influence of ionic strength on metal carbonate reaction kinetics has been investigated in a limited number of studies of calcite reaction kinetics in highly concentrated (ionic strength > 0.7 m) simple electrolyte solutions (NaCl) at above-zero temperatures (Zuddas and Mucci, 1998; Gledhill and Morse, 2006; Finneran and Morse, 2009). These studies have shown an ionic strength (salinity) dependency of the reaction rate of calcite precipitation and dissolution. Specifically, the precipitation rate of calcite has been found to increase with increasing ionic strength (at constant temperature and pCO$_2$) (Zuddas and Mucci, 1998). As outlined above, the current experimental design cannot yield information that would allow separation of the effect of ionic strength (at constant temperature) from the temperature effect (at constant ionic strength) on the reaction rate of ikaite as a function of solution saturation. The variation of pCO$_2$ conditions at constant salinity and temperature yielded limited information on the combined effect of these two physical parameters when coupled in the ice–brine system on the surface of the polar oceans. Sufficient data for statistically significant functions (at better than 90% significance level) of log($R_{SA}$) versus log($\Omega - 1$) at individual salinity–temperature settings were acquired only at

![Figure 3](image-url)
S = 65.7 and t = −3.6 °C (r² = 0.939, p = 0.004, n = 5) within 1.1 < Ω < 2.4, and at S = 101.8 and t = −5.9 °C (r² = 0.965, p = 0.085, n = 3) within 1.1 < Ω < 3.7. This analysis yielded a lower reaction order at S = 101.8 and t = −5.9 °C (n = 0.68 ± 1.14) than at S = 65.7 and t = −3.6 °C (n = 1.31 ± 0.52) but a similar reaction rate constant in both cases (log kppt = 2.01 ± 0.31 at S = 65.7 and t = −3.6 °C; log kppt = 2.19 ± 1.20 at S = 101.8 and t = −5.9 °C). The latter is evident in the convergence of both trends at log(Ω − 1) = 0 (Ω = 2) (Fig. 4); beyond this point, i.e., Ω > 2, a change of slope (i.e., reaction order) of the CaCO₃ precipitation function of logR versus log(Ω − 1) has been seen for calcite in seawater at 25 °C (Zhong and Mucci, 1993). The above suggest that, in addition to the Ωₙkaite of the brine, the salinity and temperature of the ice–brine system appear to exert control on ikaite precipitation kinetics in sea ice via a change in the reaction order in the disequilibrium region close to saturation. The fitted function on all available data reported in Section 4.1 (Fig. 3a) should nevertheless provide a good approximation of the precipitation rate of ikaite within the experimental Ωₙkaite range in sea ice brines.

The ionic strength of the solution has been found to have the reverse effect on the dissolution rate of calcite, but the drop in the dissolution rate of calcite with increasing ionic strength was not evident at pCO₂ < 0.1 atm (Gledhill and Morse, 2006; Finneran and Morse, 2009). A minimal effect of ionic strength is, therefore, expected on the dissolution rate of ikaite at the much lower pCO₂ conditions during ikaite dissolution of this study. This is implied in the comparatively tight distribution of the current observations around the linear function of log(1 − Ω) (Fig. 3b), suggesting that the saturation state of the seawater or brine with respect to ikaite and, by extension, the pCO₂ of the system should be the main controlling parameter of the dissolution rate of this polymorph.

Finally, the derivation of a rate law from bulk kinetic observations is affected by how well the stoichiometric equilibrium solubility of the solid phase is known, along with the stoichiometric equilibrium constants that describe the equilibrium reactions amongst gaseous CO₂ and aqueous CO₂ species (Hales and Emerson, 1997). The stoichiometric solubility of ikaite (Table 2) has been concurrently measured in the same solutions during long-term closed system incubations as reported in Papadimitriou et al. (2013), but the equilibrium constants that regulate the oceanic CO₂ system are not known from observation in the sub-zero hypersaline conditions of this study; only extrapolation from the fitted observations in the standard oceanic salinity and above-zero temperature range is available to date. The error incurred by the extrapolation is unknown. It is with this caveat that the empirical rate law for ikaite dissolution and precipitation in seawater-derived brines at sub-zero temperatures was obtained and was discussed here across the entire experimental temperature-salinity range.

4.3. Implications for the ikaite cycle in sea ice

With the above stipulation in mind, an attempt will be made below to estimate the time scale of the ikaite–solution reaction in the polar environment by coupling the empirical rate law with the simple model of the brine CO₂ system described in Section 2.4. In the absence of gas exchange, solid–solution reaction, and biological activity during sea ice formation and growth, the physical concentration of the seawater solutes generates a gas–liquid CO₂ concentration in the brine in a closed system leads to conservation of the surface oceanic A T and C T in the brine channels and pores of sea ice. This will result in the rise of brine pCO₂ to values more than an order of magnitude higher than the current atmospheric concentration (Table 4). In defence of this result, a commensurate trend has been observed in artificial sea ice (Geißfuss et al., 2012), and values as high as 12,000 μatm have been documented in winter sea ice (Miller et al., 2011a,b). Under these elevated pCO₂ conditions, the physical-concentration-driven rise in CO₂⁺ (Table 4) and Ca²⁺ is insufficient relative to the empirical Kᵦₙₚₖᵦₜₑ for the brine to become saturated with respect to ikaite and, thus, viable for ikaite precipitation in the temperature range from the freezing point of seawater down to −8 °C. In other words, the brine remains corrosive to ikaite (Table 4; Ωₙₖₜₑ < 1.0).

The increase in brine pCO₂ during the physical concentration of seawater solutes generates a gas–liquid CO₂ disequilibrium, a driver of potential CO₂ escape from the brine into gas bubbles. Brine degassing during wintertime cooling and inorganic carbon uptake by the sympagic autotrophs in the productive layers of sea ice in the spring–summer season are two mechanisms which can depress the brine pCO₂ (Gleitz et al., 1995; Papadimitriou et al., 2007, 2012; Delille et al. 2007; Munro et al., 2010). The lowering of brine pCO₂ will result in re-equilibration of the brine CO₂ system, which yields higher pH and CO₂⁺ concentration in the brine than the values derived from physical sea salt concentration alone (Table 4). This, in turn, will result in
条件 favourable for the precipitation of ikaite at the temperature where the solubility of the polymorph is exceeded ($\Omega_{\text{ikaite}} > 1$). For example, the brine (derived from the surface seawater composition outlined in Section 2.4) at thermal equilibrium with ice at $-4.7 \degree C$ ($S = 79.5$) will reach saturation with respect to ($\Omega_{\text{ikaite}} = 1$) after its pCO$_2$ has been lowered from the conserved value to atmospheric equilibrium (ca. 390 μm; Papadimitriou et al., 2013). When the brine cools to temperatures below this onset temperature and, thus, becomes more concentrated but retains its equilibrium with atmospheric pCO$_2$, the $\Omega_{\text{ikaite}}$ will rise further to values as high as 2.7 at $-8 \degree C$ (Table 4).

The above analysis shows that, at best, conserved $A_T$ and non-conserved $C_T$ through the lowering of brine pCO$_2$ to atmospheric equilibrium will afford modest values for $\Omega_{\text{ikaite}}$ within the sea ice system (i.e., $\Omega_{\text{ikaite}} = 1–3$) in the temperature range from $-2 \degree C$ to $-8 \degree C$ (Table 4). This extent of ikaite supersaturation in brines and, in the absence of information for the surface area of authigenic ikaite particles, the bulk solution reaction rate ($R_{\text{bulk-sln}}$, see Section 2.5 for definition; Table 3) will be used here, to compute an estimate of the time scale of ikaite formation in sea ice. The $R_{\text{bulk-sln}}$ was fitted to the same empirical function as $R_{\text{SA}}$ (see Section 4.1), yielding, $\log R_{\text{bulk-sln}} = 0.70 + 1.25 \log \Omega - 1$ for the precipitation reaction ($r^2 = 0.746$, $p < 0.001$, $n = 12$) and $\log R_{\text{bulk-sln}} = 1.16 + 1.12 \log (1 - \Omega)$ for the dissolution reaction ($r^2 = 0.865$, $p < 0.001$, $n = 16$). The steady state reaction rates from the current experiments suggest that the time scale for the ikaite precipitation reaction to reach equilibrium in sea ice is in the order of 1–3 weeks (Table 4), similar to the time frame documented as ikaite occurrence from onset of seawater freezing in young sea ice in field studies (Rysgaard et al., 2012, 2015). The observed dissolution rates imply that, in the severely corrosive conditions of polar surface oceanic seawater (see Section 2.4), ikaite will dissolve at a rate of 281 μmol kg$^{-1}$ d$^{-1}$, equivalent to 0.059 g ikaite kg$^{-1}$ d$^{-1}$. An area-weighted estimate for the sea ice ikaite content of 0.3–3.0 g m$^{-2}$ was derived from field observations in the Weddell Sea (Dieckmann et al., 2008). Using the dissolution rate above, the reported ikaite content of sea ice will dissolve in a time frame ranging from a few hours to a few days in the surface ocean, assuming that it escapes dissolution during sea ice decay and melt, and is homogeneously mixed throughout a surface mixed layer of 10–50 m thickness.

5. CONCLUSIONS

The precipitation and dissolution rates of ikaite in seawater and seawater-derived brines at sub-zero temperatures were found to conform to the same empirical rate law as those of the anhydrous CaCO$_3$ polymorphs, calcite and aragonite. The reaction order with respect to the deviation of the solution from ikaite saturation was close to first order reaction kinetics, while the reaction rate constant was higher than that of the anhydrous polymorphs. In addition to the saturation state of the brine with respect to ikaite, the salinity and temperature of the brine appear to exert control on ikaite precipitation kinetics, while the dissolution kinetics of the polymorph appear impervious to their influence.

Simple modelling of the brine CO$_2$ system demonstrated that the physical concentration of seawater solutes coupled with the changes in ikaite solubility in the temperature and salinity conditions of sea ice formation and growth to $-8 \degree C$ are insufficient to drive the internal sea ice brines to ikaite precipitation. The loss of dissolved inorganic carbon to the gas phase of sea ice and, with potentially more dramatic results, to the sympagic autotrophic community can be a potent trigger of the ikaite cycle via regulation of the brine pH and, with it, the carbonate ion concentration in the brine through the $C_T$ species re-equilibration.

Based on this, the precipitation of ikaite from sea ice brines in the cooling phase of the system (autumn–winter) will conceivably depend on the rate of temperature change.
(i.e., heat exchange in the ocean–ice–air system), which will cause, as a cascade of events, the physical concentration of solutes, the solubility changes of the polymorph, and also gas solubility changes in the sea ice brines. Added to this, the rate and extent of $C_T$ loss from the brine via CO$_2$ degassing, which is related to the gas solubility changes, and via primary production will be of primary importance as independent mechanisms of compositional modification of the sea ice brines towards ikaite supersaturation and precipitation. Finally, the observed dissolution kinetics of ikaite suggest short turn-over time scales of the polymorph in corrosive solutions, such as surface oceanic water.

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