TIMING OF METHANE RELEASE FROM HYDRATE DISSOCIATION ON THE WEST SVALBARD MARGIN

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ABSTRACT

The release of methane from methane hydrate has been invoked as a contributing agent to rapid climate change. In 2008, plumes of methane bubbles were observed emanating from the seabed in an area of the West Svalbard margin that has undergone 1°C of warming over the last 30 years. The locations of the plumes, spreading upslope from the present upper limit of the methane hydrate stability zone, indicate that the gas in the plumes could be methane from warming-induced hydrate dissociation. Through numerical modelling, we investigated the controls on the time lag between the increase in seabed temperature and gas flow from the seabed, which results from the time taken for heat diffusion, hydrate dissociation and gas migration. These are dependent on the distribution and concentration of hydrate and on sediment permeability, among other parameters. We investigated possible scenarios in which plumes could be explained by recent warming. If the top of the hydrate was initially 5 m below sea floor, in sediment with permeability $1e^{-14}$ m$^2$, the time lag would be 40 years, which is too long to explain present-day gas emission. If, however, the presence of fractures increases permeability, the lag time could be reduced to less than 30 years, and if the hydrate is shallower the lag time can be reduced greatly; to no lag at all in the extreme case of when hydrate is at the seabed at the limit of the gas hydrate stability zone. The temperature history of the area is an important controlling factor. A 60-year time series, constructed from extrapolation of more southerly oceanic temperature records, shows that prior to the last 30 years warming, the area was cooling from a warm period in the early 1960s. So we cannot appeal to a longer period of warming to explain the present-day methane release, but earlier warm periods could have brought gas to the seabed, which, during the subsequent cooling, formed hydrate close to the seabed. Dissociation of this shallow hydrate over the last 30 years would produce methane release from the seabed at the present day. Models over a 1000-year period, demonstrate the role of temporal variation in seabed temperature in bringing the hydrate close to the seabed.

Keywords: gas hydrate, methane release, ocean warming

INTRODUCTION

The release of methane to the atmosphere from methane hydrate in sediment beneath the ocean floor has been invoked as a contributing agent to rapid climate-warming events in the past, such as the Palaeocene-Eocene Thermal Maximum PETM [1] and Quaternary glacial-to-interglacial transitions [2]. A current concern is that modern climate warming will destabilise methane hydrate, releasing methane to the atmosphere. Methane is

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Hydrate is stable under conditions of low temperature and high pressure. In the Arctic, colder water at the seabed allows hydrate to form in shallower water than elsewhere. If sediment containing hydrate is warmed sufficiently, the hydrate will dissociate to produce gaseous methane and water. Hydrate in shallow water and at a shallow depth beneath the seabed will be affected by climate warming earliest and most strongly [4]. As the Arctic is predicted to experience the most intense warming over this century [5], the Arctic is particularly vulnerable to dissociation of hydrate.

Sonar images obtained from the West Svalbard margin (Fig. 1) in Autumn 2008 showed hundreds of bubble plumes emanating from the seabed, in close association with the landward outcrop of the base of the methane hydrate stability zone [6]. Water samples taken from the plumes contain concentrations of dissolved methane that are up to twenty times higher than background. These observations could be explained by the warming-induced dissociation of hydrate leading to a release of methane into the ocean. Temperature measurements from ctd casts over the 33-year period from 1975 until 2008 show a 1°C warming of the bottom water in the range 300-450 m water depth [6].

The plumes of bubbles come from seeps in the seabed in water depths shallower than 400 m, in the uppermost part of the continental slope, and extending on to the continental shelf, which is covered by glacially derived sediment (Fig. 2). A depth of 400 m is approximately the depth beneath which methane hydrate is stable in seawater at the present-day temperature of the seabed (3°C), i.e. the top of the gas hydrate stability zone (GHSZ).

In this paper we investigate the hypothesis that the methane in the plumes is derived from hydrate dissociated by ocean warming. We test the hypothesis by examining, through numerical simulations, the dependence of:

a) the time lag between warming of the seabed and the emission of gas from the seabed and
b) the rate of emission of gas upon:
   i) the distribution and concentration of the hydrate in the sediment,
   ii) the type and distribution of permeability in the sedimentary section containing the hydrate and
   iii) the input of methane into the base of the system.

We explore the effect of probable histories of seabed-temperature variation on gas release and on the distribution of hydrate in the hydrate stability zone.
Figure 2. Summary cross-section representing the uppermost part of the continental margin of western Svalbard, based on seismic and sonar data from [6, 7, 8]. The estimated present-day base of the GHSZ is shown by the white dashed line and its estimated base in 1978 is shown by the pale blue dashed line. Inset is an illustration of the processes active in the vicinity of the outcrop of the base the GHSZ at the seabed. There is abundant evidence of gas migrating up slope from source deeply beneath the GHSZ and feeding gas seeps at the edge of the shelf. The concentration of gas seeps in the area from which the GHSZ has withdrawn and their absence from the region in which the GHSZ is present indicates that dissociation of hydrate has contributed to the seeps and also that the withdrawal of the GHSZ enabled gas from beneath to migrate to the seabed through the zones in which the GHSZ existed formerly.

**MODELLING APPROACH**

Numerical modelling of fluid flow and heat flow coupled with hydrate formation and dissociation was undertaken with TOUGH+Hydrate [9]. TOUGH+Hydrate has been used previously for both 1D and 2D cases [10,11] to investigate the fate of oceanic hydrate under a warming ocean. We adopted a 1D approach, modelling different depths down the continental slope separately in a stepwise fashion (Figure 3). The dips of boundaries in the modelled zone are about 1°. The initialisation and run times of 2D models were too long to permit the multiple runs needed for sensitivity testing.

The basic 1D model (Figure 3) has a cell thickness of 0.5 m for the first 100-m depth and thereafter cell thickness increased by a factor of 1.1 for each successive cell to a maximum depth of 1134 m. The cells at the top and bottom of the model have a thickness of 0.001 m to represent the boundary conditions. There is no flow in the x and y directions. The bottom cell has a source term that is used to represent fixed heat and fluid flows. The top boundary of the model, representing the bottom of the ocean, is represented by a cell with fixed pressure, temperature and aqueous methane saturation.

The models were initialised with a hydrostatic pressure gradient and constant heat flow. The base of the hydrate stability zone was found by the code during initialisation, using the hydrate stability criterion [12] for pore water containing sodium chloride at a concentration of 35 g/kg. The values of the invariant model parameters are listed in Table 1.
The average geothermal gradient in the area was estimated at 65 °C/km from observations of the depths of the BSR in seismic reflection data. The thermal conductivity and diffusivity of water-saturated sediments were calculated for sediment porosity of 60% using the method of [13], and multiplied by the average geothermal gradient to derive heat flow. As thermal gradient changes when pore space contains hydrate rather than water, constant heat flow is required for the initial condition rather than a constant thermal gradient.

Table 1. Invariant model parameters

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Heat flow</td>
<td>7.865 x10^-2 W/m²</td>
</tr>
<tr>
<td>Water saturated thermal conductivity</td>
<td>1.21 W/m/K</td>
</tr>
<tr>
<td>Porosity</td>
<td>60%</td>
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Permeability in the model was assumed to be uniform, and different models took a range of values across seven orders of magnitude. The values of permeability given for the models are those of intrinsic (absolute) permeability of the rock, in the absence of hydrate, which reduces the permeability, depending upon its concentration. The presence of fractures introduces a dual-porosity-dual-permeability system. Where fractures predominate, the properties of the system approximate to those of the fracture network, with high permeability and low porosity, and, in cases where this was considered to be probable, layers were given a very low porosity (1%) and high permeability (10^-11 m²). The definitions of relative permeability and capillary pressure are taken from [14] and [15], respectively.

Changes in bottom-water temperature with time were applied to the model at yearly intervals by changing the temperature in the upper boundary cell. Methane flow into this cell was taken to represent methane flow into the ocean.

RESULTS

The primary elements of the response to warming are illustrated by a model of hydrate dissociation as the seabed temperature at 390 m depth warms linearly from 2 to 3 °C over a period of 33 years and then remains constant at 3°C until 100 years (Figure 4). The permeability used for the initial model is 10^-13 m². This value is higher than that which has been used by others (e.g. [11]) for modelling flow through marine clay-rich sediments. We explore the effects of varying permeability and discuss the likely values of effective permeability, below. The base of the hydrate stability field is initially at 402 m depth (12 m beneath the seabed). The top of the hydrate stability zone in the water column is initially at 373 m depth and after 15 years it reaches the seabed. Hydrate initially occupies the interval between 7 m and 12 m beneath the seabed. Gas reaches the seabed 38 years after warming starts. After 5 years, warming at the base of the hydrate stability zone produces dissociation of hydrate to methane gas and fresh water.

At 16 years, hydrate dissociation is occurring through all but the top of the layer containing hydrate and the heat energy is being used for dissociation of hydrate rather than warming deeper sediments. Gas and hydrate occur together, because as dissociation occurs, the pressure increase caused by released gas and the salinity decrease caused by released water both increase the stability of hydrate. This is compensated by the increase in temperature up to the temperature at which hydrate dissociates with decreased salinity and increased pressure. The geotherms in figure 4a remain tightly bunched close to the dissociation temperature in the region where hydrate is present, moving slightly to higher temperatures, because of the transient increase in stability, until the hydrate has dissociated. After which, the geotherms rapidly migrate towards the long-term geothermal...
gradient for the increased temperature of the seabed. The rate of dissociation is controlled primarily by the rate at which heat can be supplied, but it is also influenced by the rate at which gas can flow away from the hydrate, affecting the local pore pressure.

Gas flow is restricted by reduced permeability caused by hydrate where it is present in pore space, but permeability increases as the hydrate dissociates. Gas pressure reduces significantly and gas reaches the seabed just after the hydrate has finished dissociating at about 38 years. (Figure 4b,c,d). Thereafter, gas in pore space asymptotically declines towards its critical saturation of 3% (Figure 4d).

The controls on the length of the time lag between the onset of warming and the appearance of gas at the seabed, were explored by varying the values of parameters of the model (Figs 5-8). The effect of varying permeability was investigated, using models with different depths to the top of a 5-m-thick hydrate layer beneath the seabed at 390 m water depth, with the top of the GHSZ in the water column initially at 373 m depth (Figure 5). The time lag decreases with increasing permeability, which reduces the time for gas to flow to the seabed.

Also, the rate of hydrate dissociation is slowed by increased pressure when permeability is low and gas flow is restricted. For the case in which hydrate is at the seabed, the time lag is 15 years, which is the minimum lag, representing the time for the seabed at this water depth to move out of the hydrate stability zone, as the water becomes warmer. The minimum lag is less in shallower water, where the seabed is closer to the initial top of the GHSZ. The time lag increases as the depth of the top of the hydrate beneath the seabed increases, because of the greater distance over which heat has to diffuse downward and gas liberated from hydrate has to flow upward.

Figure 4. Results from model of hydrate dissociation as the seabed temperature at 390 m depth warms linearly from 2 to 3 °C over a period of 33 years and then remains constant at 3°C until 100 years. The permeability used for the initial model is $10^{-13}$ m$^2$. 
The amount of hydrate in the pore space has little effect on the time at which methane appears at the seabed, except when there is only a small amount of hydrate and permeability is low (Figure 6). At low concentrations of hydrate, the amount of gas released on dissociation is also very small and so the relative permeability for gas flow remains very small and gas movement through the sediments is slow. Increasing the value for critical saturation of gas from 1 to 5% increases the time lag significantly, particularly at low permeability (increases by a factor of 3.5 for a permeability of $10^{-14}$ m$^2$).

Increasing the permeability alone from $10^{-14}$ m$^2$ to $10^{-11}$ m$^2$ reduces the lag time from 42 to 29 years (Figure 6). Also, reducing the porosity alone from 60% to 5%, and thereby reducing the specific storage, reduces the lag time from 42 years to 26 years. Increased permeability and decreased specific storage typify systems in which fracture flow is significant.

The flow rate of methane gas through the seabed increases with increasing permeability, but does not continue to increase much with permeability when permeability exceeds $10^{-13}$ m$^2$ (Figure 7) The deeper the top of the hydrate, the lower the rate, because the pressure gradient is lower. For high hydrate concentration, higher flow rates are observed for low permeability (Figure 6). The critical gas saturation also affects flow rate, with low critical saturation giving a high flow rate.

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There is a maximum flow rate of about 1250 g.m\(^{-2}\) yr\(^{-1}\) across all the model runs. This maximum rate is imposed by the rate of heat input. Once at the temperature for dissociation, hydrate requires an additional 55 kJ of energy to release one mole of methane. The limiting factors for the rate of heat input are the rate of increase of the temperature of the seabed, which increases the thermal gradient, and the thermal diffusivity of the sediment. The depth of hydrate also plays a role in controlling the thermal gradient, as the temperature is tied to the dissociation temperature during hydrate dissociation, with shallower hydrate producing a steeper gradient and more rapid heat transfer. A larger quantity of hydrate produces a proportionately larger quantity of gas, but does so by producing gas over a longer period of time. This is illustrated in Figure 9, which shows how the rate of flow of gas from the seabed varies with time for different concentrations of hydrate and rates of temperature increase at the seabed. Doubling the rate of heating of the seabed from 0.03°C/yr to 0.06°C/yr cuts the lag time by about 14 years, from 42 to 28 years. The permeability used for these models is $10^{-14}$ m\(^2\).

DISCUSSION
The examples explored here are ones in which the only source of gas is hydrate and transport of methane is by gas flow or by advection in solution in water. The initial distribution of hydrate has to be defined on the basis of models derived from examples of natural occurrences of hydrate. Transport of methane by molecular diffusion is excluded, as it is in the models of [10] and [11]. For models over short periods of time, this can be justified by the much slower transport rates provided by diffusion, but when hydrate is close to the seabed and the near-seabed processes that oxidise methane, diffusion becomes important for periods of several tens of years or more. Also, the models have no source of gas, and only have validity when the rate of input of methane is small in comparison with gas production from dissociation of hydrate. The presence of gas above critical saturation, just beneath the base of the GHSZ, when added to the gas from dissociation of hydrate, increases the concentration of the migrating gas and, hence, the rate of flow of gas to the seabed. This decreases the migration time by about four years in the case of the model of Figure 4, when the underlying gas has a concentration of 5% (2% above critical saturation).

Temperature history.
We investigated the history of temperature variation to:

a) Discover whether warming earlier than 1975 had contributed to hydrate dissociation in the area,
b) See if there was a significant difference in the response of the model over the last 33 years from that produced by a linear increase in temperature over that period, and
c) Explore how changes in temperature over a one-thousand-year period preceding the 1950s could have affected the growth and dissociation of hydrate.

An analysis of oceanographic data in the northeast Atlantic from the early 1950s until 2006, taken from the transects shown in Fig. 1, [16] shows that the temperature of the water entering the Arctic in the early 1960s, although less than in the mid 2000s, was warmer than around 1980, which saw a minimum (Figure 10). Consequently, the period of recent warming is no longer than about 30 years, and a longer period of warming is not a cause of the present-day methane release.

Using the near-bottom water-temperature data from the Svalbard plume location over the period 1975-2008 [6] (Figure 10) for the temperature at the seabed, we ran a model with same properties as that discussed above and illustrated in Figure 4, except that the initial top of the hydrate lies 5 m beneath the seabed, rather than 7 m, and compared
the results with those from the same model driven by a linear increase in temperature from 2° to 3°C (Figures 11 & 12). For both models the seabed temperature remains constant at 3°C after 33 years. The results show that in major respects the model with the linear increase in temperature provides an adequate representation of the response of the system. Gas first comes to the seabed at 34 years in the linear model (Figure 11d) and at about 36 years in the data-driven model (Figure 12d). During the period 20-25 years from the start of the run (1995-2000), new hydrate forms from gas released from earlier hydrate dissociation between 3.5 and 10m in the data-driven model. Hydrate disappears completely two years later in the data-driven model than in the linear model. The large fluctuations in the geotherm in the top 5 m do not penetrate deeper, because the temperature is buffered by the enthalpy of the dissociation of hydrate.

Figure 10. Water temperature data from stations and transects in the North Atlantic (Figure 1) [16], and near-bottom temperature data from the plume area off Svalbard [6]. The series are plotted relative to their mean values and time shifted to compensate for the time taken for water to move from the latitude of southern Norway to the Fram Strait, west of Svalbard, which is about 3 years. The amplitude of the near-bottom temperature variation is less than half that of the shallower temperatures recorded by the ocean stations.

Over the 1000-year period prior to the 1950s, there has been a period of warming, the Medieval Warm Period between about 950 and 1250 AD, and a period of cooling, the Little Ice Age between the mid 15th and mid 19th centuries. These respectively show relative warming of about 0.5°C and cooling by as much as −1°C [17,18]. To provide a simple simulation of the possible effect of these long-term variations on hydrate growth and dissociation, we ran a model over a 1000-year period between 955 and 1955 AD, in which we represented the temperature by a sine wave of 1°C amplitude with a maximum in 1205 AD and a minimum in 1705 AD. The model includes diffusion and a gas source at the base. Hydrate at a concentration of 5% of pore space occupies the top 5 m of the starting model, but rapidly disappears during warming. The model is left to grow and dissociate hydrate in response to different inputs of gas at the base and the variation of temperature at the seabed.

The length of this short conference paper precludes a detailed description of the modelling. The main results are, however, that hydrate is formed within the upper few metres beneath the seabed and its presence there is sustained if the rate of supply of gas at the base is a few hundred mol.m⁻².yr⁻¹. A model with a supply of gas of 300 mol.m⁻².yr⁻¹ can build and sustain a high concentration of hydrate near to the seabed (Figure 13). The growth of this hydrate commenced around 1500 AD and after reaching a maximum around 1700 AD the near-seabed concentration of hydrate gradually reduced. We do not claim that the model represents the actual variation in a hydrate concentration over this period. The purpose of the model is to demonstrate that hydrate can be formed and sustained close to the seabed, even with loss of methane by diffusion to the seabed, if there is a supply of methane from beneath the GHSZ.

Permeability
The TOUGH+Hydrate code used for modelling assumes intergranular porosity and permeability. This simple permeability model almost certainly does not provide a complete description of the types of permeability functioning in the sediments of the upper continental margin off Svalbard. Lying at the top of the continental slope the sediments are subject to gravitationally induced down-slope extensional stress, which will promote the formation of cracks within the sediments, oriented approximately parallel to the contours of the slope, especially if processes, such as the production of gas from hydrate, produce overpressure in the sediments.
Figure 11. Results from model with initial depth of hydrate at 5 m. Seabed temperature warms linearly from 2 to 3 °C between 0 and 33 years and then remains constant at 3°C until 100 years.

Figure 12. Results from model with initial depth of hydrate at 5 m. The temperature at the seabed follows measured values from the plume location over the period 1975-2008 and thereafter remains constant at 3°C.
Figure 13. Variation with time of hydrate concentration beneath the seabed in a water depth of 390 m when seabed temperature is varied sinusoidally over a period of 1000 years, with a maximum amplitude of 3.3°C at 250 years and a minimum of 1.3°C at 750 years. The model is fed by a supply of methane of 300 mol.m$^{-2}$ yr$^{-1}$ at its base, and molecular diffusion of methane is modelled, as well as its advection in solution and flow as a gas.

A large submarine slide immediately to the north of the principal area in which the plumes of methane bubbles exist, bears witness to the effect of gravitationally induced stress at this locality [19] and, farther down slope, S-wave anisotropy indicates the presence of aligned cracks in the shallow sediment [20]. The presence of cracks increases the hydraulic diffusivity greatly, because the hydraulic conductivity of the cracks is very high, but the small volume occupied by the cracks gives the crack network low specific storage. An increase of four orders of magnitude in hydraulic diffusivity has been observed in shallow clay-rich sediments in the outer part of the Niger delta during a natural hydrofracturing event [21]. Furthermore, numerical modelling of gas invasion in sediment [22] predicts that, in fine-grained sediments, hydrate is likely to form veins in a fracture network.

In the light of this information, it is very probable that the sediments exhibit dual permeability, in which the component of permeability provided by cracks is more important in the low permeability glacigenic sediments. Permeability estimated from general empirical data for the permeability of fine-grained marine sediments is likely to be much too low for the situation modelled here and the preferential occurrence of hydrate in cracks will bias the effective permeability of the system to that of the crack network as the hydrate dissociates in the cracks. Consequently, we consider that permeability as high or higher than 10$^{-13}$ m$^2$ should be employed to represent the natural system adequately in the type of model that we have used. The presence of fractures would enable gas from dissociating hydrate deeper than 4 m beneath the seabed to reach the seabed in 390 m water depth or shallower (limited by the decreasing depth of the base of the GHSZ) in much less than 30 years.

One-dimensional approximation
The 1D approximation used in the models limits the degree to which the models represent what occurs in nature. However, we consider that the 1D models are useful for indicating the kinds of responses to a warming event that could be expected from a natural sub-seabed hydrate system. In general, 2D and 3D models have shorter response times than 1D models, because they have flows of heat and gas in more than one direction. Heat can enter a 3D body containing hydrate and the gas escape from it more rapidly than a layer in one dimension. Gas migration in nature is not a uniform process. Migrating gas forms trains of gas at a high concentration, in which the gas moves more rapidly than the same volume of gas moving uniformly through the same formation. 1D models cannot represent this behaviour. The plumes of bubble of methane gas off Svalbard come from specific sites on the seabed, with no obvious activity occurring between them. The sub-seabed gas flow at the seep sites will be greater than in the intervening areas. The flow of gas has become concentrated on its passage to the seabed. This behaviour gives the natural hydrate system a shorter response time than the 1D models employed in the work presented here. The rates of gas supply needed to sustain hydrate near the seabed that are obtained from the models would be appropriate for the zones that feed the pathways to the gas seeps.

The 1D model for a water depth of 390 m used as an illustration in this paper has a minimum lag of 15 years imposed by the time it takes from the onset of warming for the increase of temperature at this depth to remove the seabed from the GHSZ.
Shallower parts of the seabed in the area from which the GHSZ has withdrawn respond more rapidly. It is probable that gas released from dissociating hydrate beneath seabed that is not yet out of the GHSZ, migrates laterally, up slope, along more permeable layers to reinforce gas in the zone where the GHSZ has disappeared entirely, and farther, beyond the original landward limit of the GHSZ, to emerge at the seabed. Again, this is behaviour that would give the natural system a more rapid and more diverse response than that of the 1D models. We consider that the models, however, have value in providing probable upper bounds on the response time of the natural system.

CONCLUSIONS
The principal factors that increase the rapidity with which hydrate dissociation beneath the seabed will produce gas that will reach the seabed in response to warming of the seabed are:

a) decrease in depth beneath the seabed of the top of the sediment containing hydrate,
b) increase in permeability of the sediment,
c) decrease in porosity of the sediment
d) increase in rate at which the seabed is warmed,
e) decrease in time after the onset of warming that the seabed leaves the hydrate stability field,
f) presence of free gas just beneath the base of the GHSZ.

The concentration of hydrate does not affect the rapidity of response at concentrations above 7% of pore space, but only affects the length of time for which gas is emitted. At concentrations below 7%, the response becomes more strongly dependent on permeability, and with a concentration of less 2% of pore space there may be no flow, because insufficient gas is produced to exceed critical saturation for gas flow.

For hydrate shallower than a few metres beneath the seabed, it is likely that warming of bottom water in the upper continental margin of West Svalbard since the late 1970s has produced gas from dissociation that feeds gas flares from the seabed at the present day. Depending on the degree to which the sediments contain fractures, gas produced from hydrate that is deeper than 5 m can have contributed to the gas emissions.

To sustain hydrate within a few metres depth beneath the seabed against losses of methane to the seawater by diffusion over long periods, an underlying source of methane gas with an input of a few hundred mol. m$^{-2}$ yr$^{-1}$ is needed. This supply of gas from deeper in the system is evidently present in the continental margin of west Svalbard and is a prerequisite for shallow hydrate in continental margins generally. With an underlying supply of methane, cycling of the water temperature over periods of decades and centuries can build up hydrate at shallow depths beneath the seabed.

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