Potential distribution of methane hydrates in the world’s oceans

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Abstract. Estimates of the magnitudes and spatial distribution of potential oceanic methane hydrate reservoirs have been made from pressure-temperature phase relations and a plausible range of thermal gradients, sediment porosities, and pore fillings taken from published sources, based on two major theories of gas hydrate formation (1) in situ bacterial production and (2) pore fluid expulsion models. The implications of these two models on eventual atmospheric methane release, due to global warming, are briefly examined. The calculated range of methane volumes in oceanic gas hydrates is 26.4 to 139.1 \times 10^{12} \text{ m}^3, with the most likely value on the lower end of this range. The results for the bacterial model show a preferential distribution of hydrates at mid- to high latitudes, with an equatorial enhancement in the case of the fluid migration model. The latter model also generates a deeper and thicker hydrate stability zone at most latitudes than does the former. Preliminary results suggest that the hydrate distribution predicted by the fluid migration model may be more consistent with observations. However, this preliminary finding is based on a very limited sample size, and there are high uncertainties in the assumptions. The volume of methane hydrate within the uppermost 1 m of the hydrate stability zone and within 1°C-2°C of the equilibrium curve, assuming in situ bacterial generation, is 0.93-6.32 \times 10^{12} \text{ m}^3, or 0.0035-0.012% of the maximal estimated hydrate reservoir. Nevertheless this volume, if released uniformly over the next 100 years, is comparable to current CH₄ release rates for several important CH₄ sources. Corresponding CH₄ volumes calculated using the fluid migration model are nearly 2 orders of magnitude lower.

Introduction

It has been suggested that the reservoir of carbon in methane hydrates vastly exceeds most other carbon reservoirs [Kvenvolden, 1988a]. Therefore release of significant quantities of methane into the Earth’s atmosphere from the destabilization of methane hydrates due to the anticipated global warming could exacerbate the greenhouse effect.

Methane hydrates are a type of clathrate compound, which consists of a three-dimensional lattice framework of water ice with open cages, into which various gas molecules, such as methane, ethane, CO₂, or H₂S, can fit [Miller, 1974]. Methane hydrates are stable at depths below 200 m to several thousand meters and at temperatures below 10°C-15°C, depending on depth (Figure 1). They are widespread in marine sediments on the continental margins and in permafrost regions [Kvenvolden et al., 1993; Kvenvolden, 1988a, b; 1993; Kvenvolden and Cooper, 1987]. However, the exact amounts
Methane can form either by thermochemical dissociation of organic matter at high temperatures and depths, which percolate upward and accumulate in suitable structures or lithologic traps (thermogenic methane), or by low-temperature microbial methanogenesis in near-surface sediments. Methane hydrates that originate from either of these two sources can be differentiated by their δ¹⁸O isotope composition, presence or absence of C²⁴, ratio of methane to heavier hydrocarbons, and deuterium content [Cicerone and Oremland, 1988]. Methane hydrates in marine sediments are believed to be largely of microbial origin [Kvenvolden, 1988], although occurrences of thermogenic or mixed origin are also known [Brooks et al., 1985; Kvenvolden, 1993].

Estimates of the spatial and volumetric extent of marine gas hydrates are strongly constrained by theories of their mode of origin. The origin of gas hydrates by in situ microbial activity in marine sediments, as proposed by Kvenvolden and Barnard [1983], Kvenvolden and Grantz [1990], and MacDonald [1990], implies a widespread distribution in marginal basins and along continental margins, wherever high sedimentation rates have led to rapid burial and preservation of organic matter. It is generally assumed that local concentrations of methane exceed the methane saturation solubility in the pore water, allowing accumulation of free methane gas below the BSR. In this view the BSR defines an isothermal boundary separating the solid gas hydrate above from free methane gas below. The gas hydrate layer would then act as an impermeable seal.

In the thermogenic model, methane forms by thermal "cracking" of organic matter, including heavier hydrocarbons at temperatures >100°C at depth, followed by upward migration along faults, anticlinal limbs, or other structures and concentration of gas at the base of the BSR.

Recent seismic reflection analyses of the observed impedance contrast at the BSR have been interpreted as a lithologic change from sediment containing pore-filling hydrate above the boundary and normal, water-filled porosity, with little or no free gas in sediment below [Hyndman and Spence, 1992]. In the pore fluid migration model proposed by Hyndman and Davis [1992], hydrates concentrate directly above the BSR, as methane carried by upward moving pore fluids is removed upon reaching the hydrate stability field. While the methane is generated below the level of hydrate stability, it is not usually formed deep enough for thermogenic production. The ultimate source of the methane, however, could be either bacterial or thermogenic. Although the methane levels in the rising pore fluids could be below saturation (with respect to free gas), certain geological environments facilitate upward fluid migration and concentration of methane at the BSR.
(2) nonaccretionary subduction zones, and (3) areas of high sedimentation rates. All three environments are characterized by high fluid expulsion rates, either due to tectonic compression deformation (cases 1 and 2) or as a result of later sediment compaction (case 3). A consequence of this fluid migration model is that most hydrate accumulates within a relatively narrow zone above the BSR, such that the base of the hydrate stability zone forms a sharp, discrete boundary, whereas the upper limit is gradual and diffuse. The ramifications of these models, in particular the in situ microbial model and the fluid migration model on estimates of methane hydrate reservoirs will be explored in the following sections.

Earlier Estimates of Hydrate Abundance

Early summaries of the carbon reservoir as methane hydrate were made by the Potential Gas Committee [1981, Table 1]. More recently, Kvenvolden and Grantz [1990] have estimated the extent of offshore gas hydrates in the Arctic Basin. On the basis of results from seismic reflection profiles they assume that 75% of an offshore area in the Alaskan Beaufort and Chukchi Seas (around 520,000 km²) is underlain by gas hydrates. Further assuming that the average thickness of the hydrate stability zone (HSZ) is 400 m, that hydrates only occupy 10% of this area (i.e., 40 m based on well log records for the North Slope of Alaska [Collett et al., 1988]), that the average sediment porosity is 30%, and that 140 m³ methane evolves for each cubic meter of decomposed gas hydrate, the total yield of Arctic methane is 5.25 x 10¹¹m³ x 40 m x 0.3 x 140 = 8.8 x 10¹⁴ - 10¹⁵ m³ methane = 540 GT. Since the Arctic Basin represents 5% of the world's continental margins, extrapolation of this figure gives a global total of ~1.1 x 10¹⁵GT (or ~2 x 10¹⁶ m³ [Kvenvolden, 1988b]), an estimate comparable to that of Trefilnik et al. [1977, Table 1].

MacDonald [1990] estimates that in ocean sediments, assuming an average thickness of 500 m for the HSZ between 200 and 3000 m depth, an average porosity of 40%, hydrate formation restricted to sediments with an organic carbon content above 1% (this encompasses about 10% of the ocean sediments), and 10% filling of the available pore space, the total ocean sediment volume of gas hydrate (excluding the abyssal plains) is 1.25 x 10¹⁴ km³. This is equivalent to ~1.1 x 10¹⁵ GT for 90% filled gas hydrate, a figure close to the estimate of Kvenvolden [1988b].

Input Data

The modeling of the distribution of methane hydrates on a global basis requires global, gridded information on bathymetry, ocean bottom temperature, and organic carbon content as well as information on heat flow, thermal gradient, and thermal conductivities of the sediments. These data are available at varying horizontal resolutions. We use a horizontal resolution of 1° longitude x 1° latitude as a compromise between the fine resolution of the bathymetric data and the regional resolution of the thermal data. These data sets are described below.

Present-day climatology of seafloor temperature and depth were extracted from the three-dimensional climatology of temperature compiled by Levitus [1982] from all available station data, expendible bathythermograph (XBT), and mechanical bathythermograph (MBT) data available at the National Oceanographic Data Center. The data are gridded at 1° x 1° resolution for the globe. There are a maximum of 33 levels (layers) in the vertical direction, down to -5500 m, with thickness intervals every 10 m down to 30 m depth, every 25 m from 50 to 150 m, every 50 m between 150 - 300 m, every 100 m between 300 - 1500 m, every 250 m between 1500 - 3500 m, and every 500 m below 3500 m depth. The vertical resolution of the data is thus given by the thickness of the layer above the ocean bottom. Values below the seafloor are undefined. Because of the internal consistency with the ocean bottom temperature, we choose to use the ocean depth from this data set rather than from higher resolution bathymetric data, such as ETOPO5 [National Geophysical Data Center, 1988].

The coastal zone color scanner (CZCS) [Esaias et al., 1986] measures ocean pigment from photosynthetically active biomass. The concentration of phytoplankton pigments C can be related to remotely sensed radiance by an algorithm of the form [Sathyendranath and Platt, 1989]:

\[ C = A \left( \frac{L(L_1)}{L(L_2)} \right)^b \]

where \( L \) is radiance at wavelengths \( L_1 \) and \( L_2 \); \( A \) and \( B \) are empirical constants; and \( C \) is concentration of phytoplankton pigments in milligrams per cubic meter. \( L_1 = 443 \) nm and \( L_2 = 550 \) nm at low C; \( L_1 = 520 \) nm and \( L_2 = 550 \) nm at high C.

CZCS data thus provide a measure of marine chlorophyll concentration in the near-surface water which is indirectly related to primary production [Sathyendranath et al., 1991]. The qualitative similarity in global distributions of CZCS data and maps of sediment carbon (Plate 1) suggests that the CZCS data could serve as a digitized surrogate for geologically recent carbon deposition available for bacterial conversion to methane. However, quantitative translation between the remotely sensed radiance values and percent organic carbon is currently lacking. Furthermore, the CZCS samples a very short time period, which may not be representative of longer-term productivity and sedimentation patterns. Nevertheless, the CZCS map can be used as a general guide to regions of high surface organic matter (and by inference, high seafloor organic carbon). The CZCS data are mapped to yield a 512 x 512 array for the whole globe and are given as nondimensional indices with values typically between 0 and 2. We calculated the annual mean CZCS indices from the four seasonal distributions and spatially averaged the data to 1° x 1° resolution.

Marine heat flow data are obtained from the updated, digitized database compiled by K.E. Louden [Wright and Louden, 1989]. This database provides information on >8000 data points from the world’s oceans, listing latitude-longitude coordinates, ocean depth, tectonic setting, methodology of temperature and conductivity measurements, an assessment of data quality, thermal gradient, average conductivity of sediments, and heat flow. We divided the ocean into 22 representative regions (see Table 2). The averaged thermal gradients, conductivities, and heat flow were calculated for each region, using available data only from appropriate tectonic/geological settings, such as continental slopes, rises, and trenches, and these averaged values were assumed to apply to all grid cells within the region (Table 2). Data from ocean ridges and other tectonic settings unfavorable for hydrate formation were excluded from the averaging. The values thus obtained are therefore only approximate.

Calculation of Methane Reservoir in Oceans

The volume \( V \) of methane in methane hydrate is given by

\[ V = \Delta z \times A \times P \times H \times E \]  

(1)
Plate 1. Distribution of photosynthetically active biomass as given by the annual average coastal zone color scanner (CZCS) composites for oceans depths ≤ 2000 m (See text for explanation of CZCS indices)
Table 2. Average Sediment Thermal Properties for 22 Oceanic Regions

<table>
<thead>
<tr>
<th>Ocean Region</th>
<th>Coordinates</th>
<th>Heat Flow $Q$, mW m$^{-2}$</th>
<th>Conductivity $\kappa$, W m$^{-1}$ K$^{-1}$</th>
<th>Porosity at $z_0$, $P_0$</th>
<th>Thermal Gradient $G$, K km$^{-1}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>NW Atlantic</td>
<td>0N - 80N, 40W - 100W</td>
<td>47.0</td>
<td>1.00</td>
<td>0.61</td>
<td>34.31</td>
</tr>
<tr>
<td>NE Atlantic</td>
<td>0N - 80N, 15E - 40W</td>
<td>80.2</td>
<td>1.07</td>
<td>0.56</td>
<td>56.09</td>
</tr>
<tr>
<td>East Med-Black</td>
<td>30N - 50N, 15E - 60E</td>
<td>51.6</td>
<td>0.89</td>
<td>0.69</td>
<td>40.83</td>
</tr>
<tr>
<td>South Atlantic</td>
<td>80S - 0S, 20E - 65W</td>
<td>62.7</td>
<td>0.84</td>
<td>0.73</td>
<td>51.50</td>
</tr>
<tr>
<td>Indian</td>
<td>80S - 30N, 20E - 100E</td>
<td>58.0</td>
<td>0.96</td>
<td>0.64</td>
<td>43.62</td>
</tr>
<tr>
<td>Red-Aden</td>
<td>8N - 30N, 30E - 56E</td>
<td>100.0</td>
<td>0.84</td>
<td>0.73</td>
<td>82.20</td>
</tr>
<tr>
<td>Indonesia</td>
<td>11S - 5N, 100E - 125E</td>
<td>40.0</td>
<td>0.81</td>
<td>0.76</td>
<td>33.69</td>
</tr>
<tr>
<td>Australasia</td>
<td>80S - 11S, 100E - 170W</td>
<td>57.5</td>
<td>0.90</td>
<td>0.69</td>
<td>45.24</td>
</tr>
<tr>
<td>NE Pacific</td>
<td>50N - 65N, 100W - 170E</td>
<td>97.9</td>
<td>0.74</td>
<td>0.82</td>
<td>87.50</td>
</tr>
<tr>
<td>Cascadia</td>
<td>40N - 50N, 100W - 170W</td>
<td>102.0</td>
<td>0.89</td>
<td>0.69</td>
<td>80.67</td>
</tr>
<tr>
<td>East Pacific</td>
<td>20N - 40N, 100W - 170W</td>
<td>88.1</td>
<td>0.81</td>
<td>0.76</td>
<td>74.22</td>
</tr>
<tr>
<td>Cocos Plate</td>
<td>0N - 20N, 80W - 105W</td>
<td>63.0</td>
<td>0.80</td>
<td>0.76</td>
<td>53.51</td>
</tr>
<tr>
<td>SE Pacific</td>
<td>80S - 0S, 65W - 170W</td>
<td>37.8</td>
<td>0.83</td>
<td>0.74</td>
<td>31.41</td>
</tr>
<tr>
<td>New Guinea</td>
<td>11S - 5N, 125E - 170E</td>
<td>60.0</td>
<td>0.87</td>
<td>0.71</td>
<td>48.18</td>
</tr>
<tr>
<td>West Pacific</td>
<td>5N - 23N, 100E - 120E</td>
<td>87.0</td>
<td>0.76</td>
<td>0.80</td>
<td>76.48</td>
</tr>
<tr>
<td>Marginal Seas</td>
<td>35N - 50N, 125E - 141E</td>
<td>87.0</td>
<td>0.76</td>
<td>0.80</td>
<td>76.48</td>
</tr>
<tr>
<td>West Pacific</td>
<td>45N - 65N, 141E - 155E</td>
<td>87.0</td>
<td>0.76</td>
<td>0.80</td>
<td>76.48</td>
</tr>
<tr>
<td>Convergent Margins</td>
<td>5N - 23N, 120E - 170E</td>
<td>48.1</td>
<td>0.86</td>
<td>0.72</td>
<td>38.95</td>
</tr>
<tr>
<td></td>
<td>23N - 35N, 120E - 170E</td>
<td>48.1</td>
<td>0.86</td>
<td>0.72</td>
<td>38.95</td>
</tr>
<tr>
<td></td>
<td>35N - 45N, 141E - 170E</td>
<td>48.1</td>
<td>0.86</td>
<td>0.72</td>
<td>38.95</td>
</tr>
<tr>
<td></td>
<td>45N - 65N, 155E - 170E</td>
<td>48.1</td>
<td>0.86</td>
<td>0.72</td>
<td>38.95</td>
</tr>
<tr>
<td></td>
<td>65N - 80N, 15E - 100W</td>
<td>60.0</td>
<td>0.90</td>
<td>0.68</td>
<td>47.10</td>
</tr>
</tbody>
</table>

The thickness $\Delta z$ and $A$ are the calculated thickness and areal extent of the hydrate stability zone (HSZ), respectively; $P$ is mean sediment porosity; $H$ is the percent of the porosity within the HSZ actually filled by gas hydrate; and $E$ is the volumetric gas expansion factor of methane hydrate to methane upon decomposition. Plausible values and sources for each of these variables are discussed below.

The thickness $\Delta z$ of the HSZ can be calculated knowing the methane hydrate stability curve, the bathymetry, ocean bottom temperature $T_0$ and the geothermal gradient $G$. The intersection of the gradient with the stability curve gives $\Delta z$. The phase diagram (Figure 1) portrays the region of stability of methane hydrate as a function of temperature and depth (pressure). The equations governing the methane hydrate stability curve are given by [Kuustraa and Hammershaimb, 1983]:

$$Ps = \exp \left(14.717 - \frac{1886.79}{Ts}\right) \text{ for } 248 < Ts \leq 273$$

$$Ps = \exp \left(38.9803 - \frac{8533.8}{Ts}\right) \text{ for } 273 < Ts < 298$$

where $Ts$ is the stability temperature in K and $Ps$ is the stability pressure in kilopascals. The lithostatic gradient is assumed to be very close to the hydrostatic pressure gradient of 9.795 kPa/m (or = 0.1 atm/m [Kvenvolden and McMenamin, 1980]). Let $zo$ be the ocean floor temperature and $G = 15T/15z$, the thermal gradient in the ocean sediment. At a given depth $z$, methane hydrates are stable where the pressure equals $Ps$, if $T(z) = T_0 + G (z-z_0) < Ts$.

The thermal gradient $G$ is related to conductivity $\kappa$ and heat flow $Q$ as follows:

$$G = \frac{\partial T}{\partial z} = -\frac{Q}{k}$$

The heat flow $Q$ can be assumed to remain relatively constant with depth, whereas the conductivity $\kappa$ is a function of sediment porosity $P$ which decreases with depth below the ocean floor due to sediment compaction. The porosity decrease can vary linearly or non-linearly, depending on sediment composition, grain size distribution and other factors [Blatt et al., 1980].

Hyndman and Davis [1992] give a simple exponential variation of porosity with depth, derived from empirical data, characteristic of accretionary prism settings which will be used here. The porosity-depth curve from (5) lies within the range obtained in other studies [see von Huene and Scholl, 1991, Figure 6; Erikson and Pindell, 1993].

$$P = P_0 \exp \left[-\frac{(z-z_0)}{L}\right]$$

where $P_0$ is the sediment porosity at the seafloor and $L = 1500$ m.

The conductivity $\kappa_s$ is the geometric mean of its two major constituents $\kappa_w$ (water) and $\kappa_d$ (dry sediment [Woodside and Messmer, 1961]).

$$\kappa_s = \kappa_w^{P} \times \kappa_d^{(1-P)}$$

where $\kappa_w = 0.57$ W m$^{-1}$ K$^{-1}$ and $\kappa_d = 2.4$ W m$^{-1}$ K$^{-1}$.

Given the conductivities of water $\kappa_w$ and mean dry sediment $\kappa_d$ (assumed to remain constant with depth) and the measured conductivity of wet ocean floor sediment $\kappa_{w0}$ [from Wright and Louden, 1989], the value of the seafloor surface porosity $P_0$ at $z=zo$ can be calculated. Then using (5) and (6), $\kappa_d(z)$ can be determined down to $z-z_0 = 1500$ m.

Integration of (4) yields $T(z)$, the temperature at depth $z$.
The thermal gradient \( G \) over the depth interval \((z_i - z_0)\), where \( z_i = L = 1500 \text{ m} \), is given by

\[
G = \frac{(T_i - T_0)}{(z_i - z_0)}. 
\]  
(8)

Table 2 summarizes the thermal data used [after Wright and Louden, 1989] and the thermal gradient calculated for each ocean region. The thickness \( \Delta z \) of the potential HSZ is simply the depth difference between the top of the HSZ (usually the seafloor depth) and its base, at the depth where the subsurface temperature exceeds the equilibrium temperature, for a given thermal gradient.

The percent hydrate \( H \) occupying the HSZ is a major uncertainty in deriving the magnitude of the hydrate reservoir. Earlier calculations (Table 1) assumed a complete filling of all available pore space throughout the sediment column within the HSZ. However, wells logs and cores from the North Slope, Alaska, suggest that only 10% of the HSZ may actually contain hydrates [Collett et al., 1988]. Since the North Slope contains extensive petroleum and natural gas reservoirs, methane hydrate there could be more abundant than average because of a possible thermogenic contribution. A pore filling of 10% has also been inferred [Miller et al., 1991] from the volume of water released by hydrate decomposition in marine cores [Kvenvolden and Kastner, 1990] and from geophysical measurements. In situ bacterial generation could, in principle, yield hydrates evenly distributed throughout the HSZ, although observations indicate accumulation in discrete zones or strata [Collett et al., 1988]. For purposes of calculation we assume that hydrate uniformly fills 5% of pore space throughout the HSZ as a conservative lower limit and 10% as an upper bound.

In contrast, the fluid migration hypothesis suggests that the porosity directly above the BSR (around 50% by volume) is 1/3 to 1/2 filled by gas hydrate, the balance by pore water [Hyndman and Spence, 1992]. Furthermore, gas hydrate concentration (as inferred from pore fluid chemistry) falls off smoothly from the BSR toward the seafloor. For simplicity we assume that \( H = P/2 \) at \( z_{\text{HSZ}} \) the base of the HSZ, decreasing linearly to 0 at the seafloor \( z_0 \):

\[
H(z) = 0.5 P(z_{\text{HSZ}}) (z-z_0)/(z_{\text{HSZ}}-z_0) \quad z_0 < z < z_{\text{HSZ}}. 
\]  
(9)

The likely area \( A \) for in situ microbial production is constrained by the availability of organic carbon (0.5 - 1% by weight) on continental margins [Revelle, 1983; MacDonald, 1990]. Carbon-rich sediments accumulate in areas of high marine productivity and in regions of nutrient-rich upwelling, as shown by the similar distributions of high ocean floor organic carbon [Premuzic et al., 1982] and near-surface primary productivity [Berger, 1989]. Another factor is the preservation of organic matter in marine sediments. The amount of carbon reaching the seafloor depends not only on the primary productivity, but also on seafloor depth [Stein, 1991]. Preservation of organic matter is further controlled by such factors as the sedimentation rate and oxidation state. High sedimentation rates of terrigenous organic matter occur in estuaries, deltas, shallow marginal basins, and submarine fans. Sedimentation rate is more critical to organic carbon preservation in an open marine oxic than a restricted anoxic environment [Stein, 1991]. On the other hand, anoxic basins do not necessarily contain the highest organic C concentrations [Calvert, 1987]. In this study we used the digitized CZCS distribution as a guide to seafloor organic carbon concentration. Isolines of annual mean CZCS indices of 0.5 and greater (Plate 1) visually corresponded approximately to regions of \( \geq 0.5\% \) carbon in the map of ocean floor organic carbon concentrations (Figure 2) [Premuzic et al., 1982]. We therefore used the areas enclosed by CZCS cutoffs >0.5 and > 1.0 to represent two estimates of the likely area available for in situ microbial hydrate production.

The likely area \( A \) of hydrate occurrence for the fluid migration model is estimated as follows. At convergent margins the seaward extent of the accretionary prism (a narrow wedge of highly deformed sediments caught between the downward subducting ocean plate at the trench, and the relatively undeformed forearc basin on the upper slope) is defined by the trench boundaries taken from the PLATE database (University of Texas, Institute for Geophysics). Inasmuch as the upper prism boundary is not indicated on the PLATE or any other digitized data set and compilation of relevant data entails a major literature search, 1º latitude-longitude cells centered on the trench axes were used to approximate the area of the prism (Plate 2, top right). The resulting area is probably an overestimate, since prism widths may vary from 5 to \( >40 \) km and the cells include area on the seaward side as well. To compensate for this overestimate, the area was reduced by 50%. The convergent zones include nonaccumulating as well as accreting margins. However, both types are considered favorable for hydrate emplacement [Hyndman and Davis, 1992] and are associated with fluid expulsion [von Huene and Scholl, 1991]. Nevertheless, nonaccumulating convergent margins far from terrigenous sediment sources or from high-productivity regions (i.e., sources of carbon) in the middle of the Pacific Ocean were deleted (e.g., Ryukyu, Bonin, Mariana, Kermadec, New Hebrides, and MacQuerie Ridge trenches).

The probable areas from passive margins include deep Tertiary basins and particularly areas of high Neogene sedimentation rates, inferred from the distribution of large deltas, submarine fans, and marginal seas adjacent to major rivers and drainage basins. The tabulation of sediment discharge [Milliman and Meade, 1983] was used as a general guide to delineate regions of major river output. The GEBCO 1:10,000,000 series General Bathymetric Charts of the Oceans were consulted to outline areas of submarine fans (cones), canyons, and continental slopes down to 4000 m. These areas also take into consideration enhanced sedimentation on the lower continental shelf and upper slopes during glacial low sea level stands (around 120 m lower than present [Fairbanks, 1989]) and during early stages of deglaciation (Plate 2).

The final factor in the estimation of hydrate volume is the volumetric gas expansion \( E \) of methane hydrate as it dissociates into methane and water. The actual amount of methane in the gas hydrate is usually less than that given by the ideal formula, \( CH_4 \). Methane (diameter 4.36 Å) can occupy up to eight cages (minimum diameter 4.92 Å) within the methane hydrate crystal lattice (body-centered cubic; structure I; unit cell 12 Å) in the proportion of 8 molecules of \( CH_4 \) to 46 molecules of \( H_2O \) [Sloan, 1990, p.33, Miller, 1974]. In nature, not all of the possible methane sites in the crystal structure are actually occupied. Typical lattice occupancy rates range between 70% and 90% [Sloan, 1990, p.50, Kuusstra and Hammershaimb, 1983]. At 90% occupancy, 1 m³ of \( CH_4 \) yields 170 m³ of \( CH_4 \) under standard temperature and pressure (\( E \) in (1)) [Sloan, 1990, p. 50].

The product of the individual factors in (1) yields the total volume of methane in cubic meters. To convert to mass of carbon, the volume has to be multiplied by the density of \( CH_4 \) at STP (\( \rho \) = 0.717 × 10³g/cm³) and by the stoichiometric ratio of \( C:CH_4 \) (12/16).

A series of sensitivity calculations have been made to estimate the potential reservoirs of methane hydrate in the oceans for the
Figure 2. Distribution of high ocean sediment organic C concentration [after Premuzic et al., 1982].
two major models of gas clathrate genesis. For the in situ bacterial model (cases IB1 - IB4), CZCS cutoffs of 0.5 and 1.0 and regionally averaged porosities, decreasing with depth according to (5), are combined with the assumption of uniform hydrate fillings of 5% and 10% of available pore space. For the pore fluid expulsion model (case FE1) the hydrate filling $H$ decreases linearly from the base of the HSZ to the top, as discussed above. The area of the HSZ is taken, for convergent margins, as half the area within 1° of the trench boundary, while for passive margins the area includes major depositional basins and submarine fans (Plate 2).

**Spatial Distribution of Hydrates**

Table 3 presents the area, thickness, and volume of the potential hydrate stability zone for the in situ bacterial model. The total area of the potential hydrate stability zone increases by a factor of 2.4, the average thickness (weighted by the area of each 10° latitude zone) increases by 16% from 379 to 440 m, and the volume increases by a factor of 2.6, going from CZCS cutoff 1.0 to 0.5 (Table 3). The lower CZCS cutoff encompasses a broader, seaward extending zone of lesser marine chlorophyll concentration.

Methane volumes are proportional to $H$, the percent hydrate fillings, all other factors being equal. The calculated methane volumes range between 26.4 and $139.1 \times 10^{15} \text{ m}^3$ (Table 3). These volumes fall within the range of previous estimates from oceanic settings (Table 1). Taking $H$ values of 5% as more representative of large ocean areas and the higher CZCS cutoff of 1.0 as more likely to outline hydrate-rich areas, given the qualitative association between CZCS level and organic C in marine sediments (Plate 1), a "best guess" methane volume is $26.4 \times 10^{15} \text{ m}^3$ (case IB3, Table 3). The average thickness $\Delta z$ of the HSZ ranges between 379 and 440 m for the in situ bacterial case and 453 m for the pore fluid migration hypothesis, as compared with 400 m [Kvenvolden, 1988b, Kvenvolden and Grantz, 1990] and 500 m [MacDonald, 1990].
Table 3. Parameters Used for Estimation of Volume of Methane in CH₄ Hydrates

<table>
<thead>
<tr>
<th>Case</th>
<th>CZCS Cutoff</th>
<th>Area, 10¹²m²</th>
<th>Δz, m</th>
<th>G, K km⁻¹</th>
<th>H, %</th>
<th>E</th>
<th>P_o, %</th>
<th>P, %</th>
<th>V, x10¹¹m³</th>
</tr>
</thead>
<tbody>
<tr>
<td>IB1</td>
<td>0.5</td>
<td>31.7</td>
<td>440.0</td>
<td>5.4</td>
<td>5</td>
<td>170</td>
<td>72</td>
<td>46</td>
<td>69.5</td>
</tr>
<tr>
<td>IB2</td>
<td>0.5</td>
<td>31.7</td>
<td>440.0</td>
<td>5.4</td>
<td>10</td>
<td>170</td>
<td>72</td>
<td>46</td>
<td>139.1</td>
</tr>
<tr>
<td>IB3</td>
<td>1.0</td>
<td>13.3</td>
<td>379.1</td>
<td>5.4</td>
<td>5</td>
<td>170</td>
<td>72</td>
<td>46</td>
<td>26.4</td>
</tr>
<tr>
<td>IB4</td>
<td>1.0</td>
<td>13.3</td>
<td>379.1</td>
<td>5.4</td>
<td>10</td>
<td>170</td>
<td>72</td>
<td>46</td>
<td>52.9</td>
</tr>
<tr>
<td>FE1</td>
<td>-</td>
<td>23.0</td>
<td>453.4</td>
<td>5.4</td>
<td>50†</td>
<td>170</td>
<td>72</td>
<td>46</td>
<td>114.5</td>
</tr>
<tr>
<td>KG</td>
<td></td>
<td>400</td>
<td>50</td>
<td>140</td>
<td>30</td>
<td>17.6</td>
<td>(Arctic)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>M</td>
<td></td>
<td>500</td>
<td>40</td>
<td>156</td>
<td>40</td>
<td>19.5</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Except where otherwise indicated, the parameters have been averaged either vertically or vertically and over 22 oceanic regions (see Table 2).

Abbreviations are CZCS, coastal zone color scanner; Δz, calculated thickness of the hydrate stability zone (HSZ); G, thermal gradient; H, percent porosity within HSZ actually filled by hydrate; E, volumetric gas expansion factor of methane hydrate to methane upon decomposition; P_o, sediment porosity at seafloor; P, mean sediment porosity; V, volume of methane; IB, in situ bacterial model; FE pore fluid expulsion model; KG, Kvenvolden and Grantz [1990]; and M, MacDonald [1990].

*Value is spatially and vertically constant for IB1-4.
†This value is at base HSZ; it decreases linearly to 0 at top HSZ.

The volumes of methane from the above cited studies, when adjusted for the gas expansion factor of 170 used here, yield 21.4 and 21.3 x 10¹¹m³ CH₄, respectively, which is ~20% below the value for the best guess case, IB3 (Table 3). Considering the wide uncertainties in many of the parameters, the agreement is, nonetheless, reasonable.

The variation of the percent area of the HSZ, the average depth to the top of the zone, and its average thickness as functions of latitude are summarized in Table 4 for CZCS cutoffs 0.5 and 1.0, respectively. The HSZ is largely concentrated in the northern hemisphere (62-63%), with half or more of the potential area lying north of 40°N and maximum at 50°-60°N (Table 4). A secondary con-

Table 4. Latitudinal Distribution of the Area, Average Depth to Top, and Average Thickness of the Potential Hydrate Stability Zone as Estimated by the Bacterial Model for Two CZCS Cutoffs and by the Fluid Expulsion Model

<table>
<thead>
<tr>
<th>Latitude Zone, deg</th>
<th>CZCS Cutoff = 0.5</th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>A, 10¹²m²</td>
<td>D, m</td>
<td>Δz, m</td>
<td>A, 10¹²m²</td>
<td>D, m</td>
<td>Δz, m</td>
<td>A, 10¹²m²</td>
<td>D, m</td>
<td>Δz, m</td>
</tr>
<tr>
<td>80 to 90 N</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>70 to 80 N</td>
<td>1.59</td>
<td>1365</td>
<td>327</td>
<td>1.27</td>
<td>1426</td>
<td>328</td>
<td>0.42</td>
<td>621</td>
<td>150</td>
</tr>
<tr>
<td>60 to 70 N</td>
<td>2.34</td>
<td>1460</td>
<td>312</td>
<td>1.80</td>
<td>1382</td>
<td>298</td>
<td>1.01</td>
<td>1755</td>
<td>350</td>
</tr>
<tr>
<td>50 to 60 N</td>
<td>6.83</td>
<td>2572</td>
<td>370</td>
<td>3.19</td>
<td>2390</td>
<td>344</td>
<td>4.07</td>
<td>2533</td>
<td>367</td>
</tr>
<tr>
<td>40 to 50 N</td>
<td>5.20</td>
<td>3396</td>
<td>479</td>
<td>0.95</td>
<td>2157</td>
<td>338</td>
<td>1.11</td>
<td>2798</td>
<td>436</td>
</tr>
<tr>
<td>30 to 40 N</td>
<td>0.86</td>
<td>1652</td>
<td>256</td>
<td>0.12</td>
<td>987</td>
<td>126</td>
<td>1.08</td>
<td>2963</td>
<td>520</td>
</tr>
<tr>
<td>20 to 30 N</td>
<td>0.77</td>
<td>1466</td>
<td>269</td>
<td>0.38</td>
<td>1296</td>
<td>254</td>
<td>1.37</td>
<td>2036</td>
<td>437</td>
</tr>
<tr>
<td>10 to 20 N</td>
<td>1.37</td>
<td>2381</td>
<td>368</td>
<td>0.43</td>
<td>1775</td>
<td>263</td>
<td>4.19</td>
<td>2710</td>
<td>475</td>
</tr>
<tr>
<td>0 to 10 N</td>
<td>0.79</td>
<td>2020</td>
<td>332</td>
<td>0.23</td>
<td>1149</td>
<td>193</td>
<td>2.99</td>
<td>2802</td>
<td>503</td>
</tr>
<tr>
<td>10 to 0 S</td>
<td>0.74</td>
<td>2568</td>
<td>422</td>
<td>0.35</td>
<td>1737</td>
<td>284</td>
<td>2.27</td>
<td>2929</td>
<td>504</td>
</tr>
<tr>
<td>20 to 10 S</td>
<td>0.33</td>
<td>2393</td>
<td>514</td>
<td>0.12</td>
<td>1071</td>
<td>229</td>
<td>0.76</td>
<td>3125</td>
<td>680</td>
</tr>
<tr>
<td>30 to 20 S</td>
<td>0.14</td>
<td>1085</td>
<td>235</td>
<td>0.01</td>
<td>175</td>
<td>34</td>
<td>0.98</td>
<td>2706</td>
<td>489</td>
</tr>
<tr>
<td>40 to 30 S</td>
<td>0.24</td>
<td>1311</td>
<td>273</td>
<td>0.01</td>
<td>60</td>
<td>1</td>
<td>1.34</td>
<td>3410</td>
<td>533</td>
</tr>
<tr>
<td>50 to 40 S</td>
<td>2.11</td>
<td>3062</td>
<td>553</td>
<td>0.34</td>
<td>2003</td>
<td>365</td>
<td>0.73</td>
<td>2731</td>
<td>521</td>
</tr>
<tr>
<td>60 to 50 S</td>
<td>2.59</td>
<td>3274</td>
<td>561</td>
<td>0.92</td>
<td>3012</td>
<td>540</td>
<td>0.08</td>
<td>1633</td>
<td>524</td>
</tr>
<tr>
<td>70 to 60 S</td>
<td>4.60</td>
<td>3063</td>
<td>624</td>
<td>2.33</td>
<td>2794</td>
<td>588</td>
<td>0.34</td>
<td>809</td>
<td>173</td>
</tr>
<tr>
<td>80 to 70 S</td>
<td>1.17</td>
<td>1218</td>
<td>339</td>
<td>0.82</td>
<td>1184</td>
<td>333</td>
<td>0.31</td>
<td>515</td>
<td>201</td>
</tr>
<tr>
<td>90 to 80 S</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
</tbody>
</table>

Global total 31.64 13.28 23.03

Abbreviations are A, areal extent of the hydrate stability zone (HSZ); D, average depth to the top of the HSZ; and Δz, average thickness of the HSZ.
bacterial production implies local generation of methane within paper predict considerably different spatial distributions. In situ tion model provide a better match to observed values for over 50% the in situ bacterial model. Neither model predicts the thickness of 70.5% of the sample area as compared with only 26.2-47.0% for distributed over 42 sites (Plate 2, bottom left). The fluid migration model predicts the occurrence of the hydrate stability zone over South China Sea, and the Ross Sea, Antarctica (Plate 2, top right). Newfoundland, off eastern Greenland, off Norway, the upper con- fluid expulsion model for the Beaufort Sea, the Aleutian Basin, off of the upper continental slope, Antarctica (Plate 2, top left). Greenland Sea, off Novaya Zemlya, the Sea of Okhotsk, and much Alaska and Canada, the Aleutian Basin, Baffin Bay and the Masses and hence of continental margins in the northern hemisphere, and secondly, greater levels of photosynthetic pigments around upwelling zones at high latitudes. The average depth to the top of the HSZ (usually equivalent to the seafloor depth) is, on the whole, lower for CZCS cutoff 1.0 than 0.5 (Table 4) for equivalent latitude bands, because the region of higher marine chlorophyll levels (corresponding to CZCS cutoff 1.0; see Plate 1) occurs closer to the shore at shallower water depths. However, no systematic latitudinal trends are apparent. Except at very high latitudes, the average thickness of the HSZ at equivalent latitude bands is also consistently less for the higher CZCS cutoff, again reflecting lower average water depths, higher average 76g, and narrower HSZ (Figure 1 and Plate 1). While the average thickness of the HSZ is somewhat greater at higher latitudes for CZCS cutoff 1.0, no systematic latitudinal dependence for CZCS cutoff 0.5 is observed.

The area, thickness, and volume of the potential hydrate stability zone for the pore fluid migration model (case FE1) are listed in Tables 3 and 4. The area of the potential hydrate stability zone is 23.0 x 10^4 m^2. The average thickness of the HSZ is greater, and the area is intermediate between those calculated assuming in situ bacterial conditions. On the other hand, the CH_4 volume (case FE1) lies closer to cases IB1 and IB2, which employ the less areally re- strictive CZCS cutoff (0.5). In contrast to the in situ bacterial model, nearly half of the area of the potential hydrate zone for the fluid migration model lies within 20° north and south of the equator, with a secondary concentration between 50° and 60°N (Table 4). Also in contrast to the former model, the average thickness and depth to the top of the HSZ are distinctively smaller at high latitudes. However, except at high latitudes, the zonal thickness and depth to the top of the HSZ are consistently higher for the fluid migration than the in situ bacterial model.

Areas where hydrates theoretically can occur at relatively shallow depths are shown in red and pink in Plate 2. The in situ bacte- rial model includes sites in the Beaufort Sea off the North Slope of Alaska and Canada, the Aleutian Basin, Baffin Bay and the Greenland Sea, off Novaya Zemlya, the Sea of Okhotsk, and much of the upper continental slope, Antarctica (Plate 2, top left).

Relatively shallow occurrences of hydrates are predicted by the fluid expulsion model for the Beaufort Sea, the Aleutian Basin, off Newfoundland, off eastern Greenland, off Norway, the upper continental slope of the East Siberian and Chukchi Seas, Sea of Okhotsk, South China Sea, and the Ross Sea, Antarctica (Plate 2, top right).

Reported occurrences of known and inferred oceanic hydrates from the continental margins have been compared with calculated distributions based on the two models. There are 66 sample areas distributed over 42 sites (Plate 2, bottom left). The fluid migration model predicts the occurrence of the hydrate stability zone over 70.5% of the sample area as compared with only 26.2-47.0% for the in situ bacterial model. Neither model predicts the thickness of the HSZ accurately; yet thicknesses determined by the fluid migration model provide a better match to observed values for over 50% more of the sample areas than those of the in situ bacterial model.

Discussion

The two major theories of hydrate formation considered in this paper predict considerably different spatial distributions. In situ bacterial production implies local generation of methane within carbon-rich sediments from areas of high marine productivity and accumulation of organic matter. Hydrates can form anywhere vertically, within the stability field, with the upper limit generally corre- sponding to the seafloor. On the other hand, the fluid expulsion model proposes that hydrates form under more geologically re- stricted conditions characterized by high rates of fluid transport and also expulsion of pore fluids from sediments that were underconsolidated due to rapid burial. Methane carried in solution from below (either biogenic, thermogenic, or a mixture) will form a discrete layer upon entering the base of the HSZ, grading upward into a diffuse zone [Hyndman and Davis, 1992].

Assuming an in situ bacterial origin of oceanic methane hydrates, the methane volume ranges between 26.4 and 139.1 x 10^8 m^3, with a more likely value toward the lower end of this range. The meth- ane volume, assuming pore fluid migration, is 114.5 x 10^8 m^3, which falls within the bacterial range. Except at higher latitudes, the average depth to the top of the HSZ and thickness of the HSZ are greater for the fluid migration case than for in situ bacterial cases. The fluid-migration model provides a somewhat better match of spatial distribution and average thickness of the HSZ with known or in- ferred hydrate occurrences than does the bacterial model. How- ever, the sample volume represents only a few percent of the total estimated volume of the potential hydrate stability zone, and thus these preliminary findings may not be globally valid.

The areal extent of geologically favorable sites for the pore fluid migration model may have been underestimated, particularly around Antarctica. Although Antarctica lacks rivers (a criterion used to infer probable offshore hydrate-rich areas at passive margins; see discussion on calculation of methane reservoirs above), consider- able quantities of rock debris could have been supplied to the con- tinental shelves and upper slopes during periods of glacial retreat. Furthermore, the extensive upwelling and high levels of photosyn- thetically active biomass detected in the CZCS data around Antarc- tica (Plate 1) suggests the likelihood of rapid deposition of organic-rich sediments.

Best guess International Panel on Climate Change projections of climate change anticipate a global average 2.5°C rise in air tem- peratures and a corresponding 0.48 m rise in sea level between 1990 and 2100 [Wigley and Raper, 1992]. As the increased surface air (and sea) warming is transported toward the seafloor, any hydrates near the ocean bottom already close to the stability limits will be more vulnerable to decomposition within a relatively short time frame than deeper-buried hydrates, for which the warming pulse may take centuries to millenia to reach [MacDonald, 1990]. A more rigorous calculation of ocean warming, using a coupled ocean atmospheric general circulation model and subsurface thermal propagation, will be presented elsewhere. Here a simple calculation is made, using the areal and depth distributions of hydrates given by both models to estimate the volume of methane that may be sensitive to climate warming or sea level change. We take a 1-m thick layer at the seafloor, which may be the first layer to experi- ence any ocean bottom warming and determine the area of hydrates whose top temperature is within 2°C of the equilibrium tempera- ture or whose top is within 2 m of the equilibrium depth. For the bacterial model (CZCS = 1.0) we obtain A = 0.15 and 0.52 x 10^12 m^2, for a 1° and 2°C warming, respectively, at the ocean bottom. Setting ΔT=1 m, P=0.72 (seafloor average of 22 regions, Table 2), H= 0.05 and 0.1, E=170, (1) yields volumes of 0.93-1.86 x 10^12 m^3 and 3.16 - 6.32 x 10^12 m^3, for a 1° and 2°C warming, respec- tively (Table 5). If the corresponding masses of CH_4 (0.67-1.34 and 2.27-4.53 x 10^9 g, for the 1° and 2°C cases, respectively) were
to be released at a uniform rate over a 100-year period, this would be equivalent to annual CH₄ release rates of 6.7 - 45.3 x 10¹² g/yr. These values, although comparable to those from a number of important present-day CH₄ sources [see Cicerone and Oremland, 1988, Table 4], probably represent upper limits, in that dissolution of CH₄ in seawater and oxidation may reduce the total volumes ultimately reaching the atmosphere. These effects have been omitted here.

By contrast, the CH₄ volumes within 2 m of the equilibrium depths are significantly less (Table 5). This implies that the effect of the increase in pressure due to the projected sea level rise (only ~0.5 m, as compared to the assumption of 1-2 m used here) will likely be much less than that of the temperature rise and will probably not greatly offset the anticipated thermal destabilization.

Table 5 also suggests significantly lower volumes of methane vulnerability to destabilization, under assumptions of the fluid migration hypothesis. We estimate a volume of 0.014 and 0.047 x 10¹² m³ for a 1°C and 2°C warming, respectively, and V=0 for a 2 m sea level rise. These values are lower than the corresponding ones estimated with the bacterial model. The difference arises for two principal reasons. Firstly, even though the total hydrate area estimated using the fluid migration model is larger than that for the in situ bacterial model, the hydrates in the former case generally occur at greater depths and hence colder bottom temperatures (see Table 4). Thus a 2°C warming or a 2 m sea level rise will probably not be sufficient to affect the stability of the hydrates. Secondly, the assumed depth distribution of the hydrate pore filling yields an average H = 0.0011 in the uppermost 1 m for a globally averaged hydrate zone thickness of 453.4 m (Table 3) as compared to the vertically constant 0.05-0.10 pore fillings used in the bacterial model. Concentration of hydrates at the base of the HSZ, as implied by the fluid migration hypothesis, would therefore probably inhibit significant methane release from oceanic sediments within the next 100 years.

On shorter timescales, however, hydrates could be destabilized mechanically by massive slumping and liquefaction. Submarine slides may have been more frequent during periods of glacial sea level fall (due to pressure drop destabilization) [McIver, 1981; Nisbet, 1992; W.P. Dillon private communication, 1993]. Alternatively, rapid sedimentation may foster development of excess pore fluid pressures that lead to a reduction in sediment strength and promote instability, even on relatively low slopes [Coleman and Prior, 1988]. These conditions also favor organic carbon preservation in marine sediments, conducive to the in situ bacterial activity that generates methane. Thus the close association of methane hydrates with submarine mass movements is not fortuitous. A close association is noted between many known or inferred hydrate occurrences and mass movements [Nisbet, 1992; McIver, 1981; Kvenvolden, 1993]. The mechanical disruption of previously hydroytated sediments by the mass movement could release larger quantities of methane than would be predicted from thermal dissociation alone.

The results presented here are preliminary in nature. A number of simplifying assumptions have been made. For example, use of a regionally averaged thermal gradient or porosity may not be representative of individual sites, because of the high spatial variability of thermal data, even for similar geologic/hsstonic settings within large geographic regions. This use may lead to errors in estimating thickness of the HSZ, as suggested by the high scatter between observed and calculated Δz. Also, the correspondence between CZCS satellite data and organic carbon in ocean sediments, used here as a first approximation, should be defined more quantitatively.

Table 5. Volume of Methane Vulnerable to Climate Warming as Estimated by the In Situ Bacterial and Fluid Expulsion Models

<table>
<thead>
<tr>
<th>Warming at Ocean Bottom</th>
<th>Sea Level Rise</th>
</tr>
</thead>
<tbody>
<tr>
<td>Model</td>
<td>1°C</td>
</tr>
<tr>
<td>In situ bacterial</td>
<td>0.93</td>
</tr>
<tr>
<td>(CZCS=1.0, H=5%)</td>
<td>1.86</td>
</tr>
<tr>
<td>(CZCS=1.0, H=10%)</td>
<td>0.014</td>
</tr>
</tbody>
</table>

Units are x10¹² m³. CZCS is coastal zone color scanner; H, percent porosity within hydrate stability zone actually filled by gas hydrates.

Conclusions

The spatial and volumetric extent of marine gas hydrates have been estimated from two models of hydrate formation (1) in situ bacterial production and (2) pore fluid migration. Calculated methane volumes range between 26.4 and 139.1 x 10⁶ m³, which lie within the range of previous estimates from oceanic settings (Table 1) but are higher than the recent estimates by Kvenvolden [1988b] and MacDonald [1990, Table 3]. Our best guess methane volume is 26.4 x 10⁶ m³, at the low end of the calculated range. The two models predict different spatial distributions of hydrates. The bac-
terial model suggests mid to high latitude concentrations, whereas the fluid migration model shows an equatorial enhancement (Table 4). However, except at high latitudes, the thickness and depth to the top of the HSZ are persistently greater for the latter model. Although the fluid migration model predictions appear to be somewhat more consistent with observed and inferred hydrate occurrences than the bacterial model (Plate 2), this result is still tentative, owing to the small sample area and assumptions which cannot be tested with the data presently available.

The potential sensitivity to projected climate change has been explored by estimating methane volumes contained within the uppermost 1 m of the HSZ that lie within 2°C and 2 m of the equilibrium curve (Table 5). Uniform release of this methane (according to the bacterial model) over the next 100 years (6.7 - 45.3 × 10^{14} g/yr) would be comparable to current annual release rates from a number of important CH_4 sources [Cicerone and Oremland, 1988]. However, the methane volumes calculated using bacterial model assumptions are ~2 orders of magnitude higher than those for pore fluid migration (Table 5). This difference is attributed to (1) the greater depths of occurrence (Table 4), hence colder bottom temperatures and (2) the assumption of a linear decrease in hydrate filling up the stratigraphic column, in the case of the latter model. Clearly, while disregarding for the moment the downward conduction into the marine sediments of any thermal anomaly caused by oceanic warming, the two models as outlined here differ vastly in the predicted volumes of methane that could be potentially released in the foreseeable future owing to minor superficial warming of the seafloor. The pressure stabilization, due to the projected sea level rise of -0.5 m, is expected to be significantly less important than the thermal destabilization (Table 5). An important test of the hypotheses proposed here will depend on the actual vertical distribution of hydrates within the ocean sediment column. Finally, even minor methane release due to thermal dissociation of hydrates at the seafloor could be vastly amplified, owing to the triggering of numerous submarine slides.

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