

Mathematical simulation of the methane discharge into the atmosphere caused by decomposition of the methane hydrates of the World Ocean

A.V. Shcherbakov and V.V. Malakhova

*Institute of Computational Mathematics and Mathematical Geophysics, Novosibirsk
Yugorskii Research and Development Institute of Information Technologies, Khanty-Mansiisk*

Received November 30, 2004

We have investigated decomposition of methane hydrates in the World Ocean that may occur due to the secular change in the surface temperature of the ocean. Several scenarios of warming or cooling of the surface water are considered. It is supposed that the temperature of the ocean surface changes by 1 to 1.5 degree for 50 years. The climatic model of the ocean thermohydrodynamics, complemented with the heat conduction equation for the bottom sediment and the transport equation for dissolved methane, is integrated over a period of 2000 to 8000 years. The transport of dissolved methane from methane hydrates buried under the bottom at a depth from 100 to 300 m is simulated. The obtained methane flows into the atmosphere, comparable with the available estimates, vary from 1.0 to 10 Tg/year, depending on the warming scenarios. The time of beginning of possible decomposition of the methane hydrates varies from 70 to 1 000 years depending on the scenario parameters.

According to modern ideas on the mechanisms of global warming of the atmosphere due to the greenhouse effect, methane is responsible for 15% of the increase in the mean temperature. Although being fourfold lower than analogous estimate for carbon dioxide (61%), the potential contribution of methane to the climate change is 21 times higher than that of CO₂. Gas hydrates may contain three orders of magnitude greater amounts of methane than the amount of methane in the nowadays atmosphere. Since disturbances of the hydrate sediments, resulting in gas leakages into the atmosphere, may cause the increase of the global temperature, gas hydrates may play certain role in the global climate change. The expected climate warming in the 21st century may lead to the shift of the phase boundaries, decomposition of gas-hydrate forms, and anomalous methane emissions into the atmosphere, favoring the enhancement of the greenhouse effect.

In our previous papers,^{1–3} we considered possible consequences of destabilization of the methane hydrates, which contact with the bottom water, under the effect of increase of the surface ocean temperature. It was assumed that methane hydrates exist on the bottom and the continental slope everywhere, where thermobaric conditions are favorable for their existence. As soon as the heat from the ocean surface achieves the bottom water and increases its temperature by a certain value — tolerance,⁴ the destabilization of methane hydrates begins. In the model, the destabilization is defined as the increase of the methane concentration at a near-bottom point.

It turned out that under these assumptions the period from the beginning of warming of the surface water to the beginning of possible destabilization is

not long, about several years, because in the high latitudes the heat from the surface water quickly reaches the bottom water due to convective mixing. The possible destabilization of gas hydrates in different scenario experiments gives the methane flows into the atmosphere about 3–14.5 Tg/yr.

The data on distribution of the ocean methane hydrates indicate that methane hydrate usually lie under the bottom at the depth of about hundreds of meters. Therefore, for methane hydrates to decompose, the heat should penetrate not only to the bottom, but also inside the sediment layer, which takes longer time as compared to the case of methane hydrates, which are in the immediate contact with water. The possible decomposition of methane hydrates can be estimated (both in the amount and in time) using mathematical models, which describe the heat redistribution in the system "atmosphere–ocean–sediment layer of the sea bottom," based on the information about the distribution of the methane hydrates.

Such an attempt is presented in this paper. The 3D quasi-geostrophic model of the World Ocean dynamics is combined with the model of heat transfer in the bottom sediment layer to study the process of propagation of a thermal signal deep into the ocean and into the sediment layer. The same model, complemented with the transport equation for dissolved methane, is used in scenario numerical experiments to estimate the amount of methane emitted into the atmosphere.

The data on the paleoclimate⁵ show that climatic changes are periodic. Warming alternates cooling, and the periods of these oscillations are from tens to thousands years. Therefore, this paper considers the scenarios of both warming and cooling of the surface water with a period of 100 years.

The climatic state of the ocean was obtained from the 3D model of the World Ocean dynamics, including the seasonal variability, with regard for the actual topography of the bottom.⁶ The problem was solved in the polygonal area of the World Ocean on a 5-degree latitude–longitude grid from 72.5°S to 87.5°N with 24 vertical levels until the achievement of the quasi-stationary state in the period of several thousands of years.⁷

The equations describing the ocean climate have the form

$$R_1 u + \ell v = -\frac{1}{\rho_0 \sin \theta} \frac{\partial P}{\partial \lambda} + \frac{\partial}{\partial z} v \frac{\partial u}{\partial z}; \quad (1)$$

$$-\ell u + R_1 v = -\frac{1}{\rho_0} \frac{\partial P}{\partial \theta} + \frac{\partial}{\partial z} v \frac{\partial v}{\partial z}; \quad (2)$$

$$\frac{1}{a \sin \theta} \left(\frac{\partial u}{\partial \lambda} + \frac{\partial v \sin \theta}{\partial \theta} \right) + \frac{\partial w}{\partial z} = 0; \quad (3)$$

$$P = -g \rho_0 \zeta + g \int_0^z \rho dz; \quad (4)$$

$$\frac{\partial T}{\partial t} + \frac{u}{a \sin \theta} \frac{\partial T}{\partial \lambda} + \frac{v}{a} \frac{\partial T}{\partial \theta} + w \frac{\partial T}{\partial z} = \frac{\partial}{\partial z} k \frac{\partial T}{\partial z} + \frac{\mu}{a^2} \Delta T; \quad (5)$$

$$\frac{\partial S}{\partial t} + \frac{u}{a \sin \theta} \frac{\partial S}{\partial \lambda} + \frac{v}{a} \frac{\partial S}{\partial \theta} + w \frac{\partial S}{\partial z} = \frac{\partial}{\partial z} k \frac{\partial S}{\partial z} + \frac{\mu}{a^2} \Delta S; \quad (6)$$

$$\rho = \rho_0 + 10^{-3} [0.802(S-35) - T(0.0735 + 0.00469T)]. \quad (7)$$

The boundary conditions are:

at $z = 0$:

$$v \frac{\partial u}{\partial z} = -\frac{\tau_\lambda}{\rho_0}, \quad v \frac{\partial v}{\partial z} = -\frac{\tau_\theta}{\rho_0};$$

$$w = 0; \quad T = T^*(t, \lambda, \theta, 0); \quad S = S^*(t, \lambda, \theta, 0); \quad (8)$$

at $z = H$:

$$v \frac{\partial u}{\partial z} = -R_2 \int_0^H u dz, \quad v \frac{\partial v}{\partial z} = -R_2 \int_0^H v dz;$$

$$w = \frac{u}{a \sin \theta} \frac{\partial H}{\partial \lambda} + \frac{v}{a} \frac{\partial H}{\partial \theta}; \quad k \frac{\partial T}{\partial z} = 0, \quad k \frac{\partial S}{\partial z} = 0. \quad (9)$$

At the lateral surface Γ :

$$\mu \frac{\partial T}{\partial n} = 0; \quad \frac{\partial S}{\partial n} = 0; \quad u_n = 0. \quad (10)$$

At the initial time $t = 0$:

$$T = T^*(\lambda, \theta, z), \quad S = S^*(\lambda, \theta, z). \quad (11)$$

The following designations are used: u, v, w are the velocity vector components along the directions λ, θ, z , where λ is the longitude, θ is latitude, and the z axis is directed vertically downward; t is time; P is pressure; ρ_0, ρ are the mean value and the anomaly of density; $\zeta = \xi - P_{\text{atm}}/(g\rho_0)$ is the reduced level; P_{atm} is the atmospheric pressure; $z = \xi(\lambda, \theta)$ is

the equation of the ocean surface; $R_1 u, R_1 v$ is the parameterization of the horizontal turbulent viscosity; v is the vertical turbulent viscosity coefficient; k and μ are the coefficients of the vertical and horizontal turbulent diffusion of heat, salt, and dissolved methane; $\ell = 2\omega \cos \theta$ is the Coriolis parameter; a, ω , and g are the Earth's radius, angular velocity, and acceleration due to gravity;

$$\tau_\lambda(t, \lambda, \theta), \tau_\theta(t, \lambda, \theta)$$

are known wind friction stresses;

$$T^*(\lambda, \theta, z), S^*(\lambda, \theta, z)$$

are the calculated 3D climatic fields of temperature and salinity, which were obtained by integrating the model (1)–(7) until the achievement of the quasi-stationary state;

$$T^*(t, \lambda, \theta, 0), S^*(t, \lambda, \theta, 0)$$

are the seasonally varying surface temperature and salinity, defined on the ocean surface; R_2 is the bottom friction coefficient; n is the normal to the surface Γ ; $H(\lambda, \theta)$ is the relief of the ocean bottom.

The ocean model is complemented with the 1D model of the bottom sediment layer having the thickness $d = 1000$ m:

$$\frac{\partial T_s}{\partial t} = \frac{\partial}{\partial z} k_s \frac{\partial T_s}{\partial z}; \quad (12)$$

$$z = H: T_s = T_{s,B}, \quad (13)$$

$$z = H + d: k_s \frac{\partial T_s}{\partial z} = Q_t; \quad (14)$$

$$t = 0: T_s = T_{s,C}(z), \quad (15)$$

where $k_s = 10^{-2}$ cm/s is the thermal diffusivity of the sediment layer; $Q_t = 3^\circ/100$ m is the geothermal heat flux; $T_{s,C}(z)$ is the solution of the stationary equation (12) with the climatic temperature $T_{s,B}$ from the problem (1)–(10).

The transport of dissolved methane is described by the equation

$$\frac{\partial C}{\partial t} + \frac{u}{a \sin \theta} \frac{\partial C}{\partial \lambda} + \frac{v}{a} \frac{\partial C}{\partial \theta} + w \frac{\partial C}{\partial z} = \frac{\partial}{\partial z} k \frac{\partial C}{\partial z} + \frac{\mu}{a^2} \Delta C \quad (16)$$

with the boundary conditions:

$$z = 0: C = C^*(\lambda, \theta, 0),$$

$$z = H: k \frac{\partial C}{\partial z} = 0; \quad (17)$$

$$\text{at } \Gamma: \mu \frac{\partial C}{\partial n} = 0.$$

Here C is the concentration of dissolved methane, $C^*(\lambda, \theta, 0)$ are the preset values of the methane concentrations on the ocean surface, which change zonally from 50 ppb in the southern high-latitude ocean areas to 100 ppb in the northern ones.

Several scenario experiments have been conducted. Each of the experiments started using the same initial climatic fields calculated. It was assumed, as before, that methane hydrates exist everywhere in the sediment layer, where the thermobaric conditions for their existence are fulfilled. In all the experiments, the time to the beginning of destabilization of the methane hydrates was estimated, the bottom part containing such sources was determined, and the diffusion methane flow into the atmosphere was found.

In the "Warming" experiment, the seasonally varying temperature of the ocean surface everywhere, except for polar regions, which are covered with ice whole year round, first increased linearly by 1.5° for the first 50 years of the experiment and then decreased also linearly by 1.5° to the initial values for 50 years as well. Then the temperature of the ocean surface continued to vary in accordance with the seasonal climatic data. The thermal signal from the ocean surface was transported into the ocean depth to near-bottom layers by convective mixing and by currents. Then, using the one-dimensional thermal conductivity equation (16) with the conditions (17), the increase in the temperature of the sediment layer as compared to the initial climatic values was simulated. As soon as this increase exceeded some acceptable value (tolerance⁴), which was a scenario parameter, for example, $dT_h = 0.01^\circ$, and this occurred at a certain depth below the bottom, for example, $H_h = 100$ m, at which the methane hydrates are buried and which was also a scenario parameter, it was assumed that decomposition of the methane hydrates begins.

This decomposition maintains the near-bottom concentration of dissolved methane equal to 5000 ppb. It is believed that methane, released upon the decomposition, fully dissolves in the seawater. The obtained value is then used as a boundary condition at this point of the bottom for the solution of the methane transport equation in place of the boundary condition with the zero methane flow, which was used to obtain the climatic data. Thus, a methane source appears in the model. If in the process of experiment the heat increment in the sediment layer

at the depth of methane hydrates becomes smaller than the preset parameter dT_h then the destabilization is assumed terminated and the source "turns off." This means that the boundary condition for the methane transport equation again changes from the preset value of 5000 ppb to the condition with the zero methane flow. This approach is confirmed by the measured data on the methane concentration in the ocean water.

Thus, according to measurements in the Sea of Okhotsk,⁸ where methane gas hydrates were found, the methane concentration near the gas flow from the decomposing methane hydrates appeared to be equal to 20000 ppb.

In the "Warming–Cooling" experiment, the seasonally varying temperature of the ocean surface changed for the first 100 years as in the "Warming" experiment, and in the next 100 years the cooling occurred. In this case, the surface temperature first decreased at a rate of 1.5° per 50 years and then again increased with the same rate to the initial seasonal climatic values, and then the ocean surface temperature continued to change in accordance with the seasonal climatic data.

It turned out that the model of the global ocean climate stronger and longer reacts to the cooling of the surface water than to the warming.⁹ Thus, in the "Warming" experiment, it took about 700 years for the ocean climate model to return into the initial state, while in the "Warming–cooling" experiment it took much longer – about 1700 years.

The methane model with the bottom sediment layer reacts in the same manner (Fig. 1 and the Table). The time from the beginning of destabilization in both of the experiments is 76 years with the temperature tolerance $dT_h = 0.01^\circ$, the methane hydrate depth $H_h = 100$ m in the Southern and Arctic Oceans and $H_h = 200$ m in other oceans. The methane flow into the atmosphere achieved its maximum value of 2.5 Tg/yr in 438 years after the beginning of the "warming" experiment and 9.1 Tg/yr in 1715 years after the beginning of the "Warming–Cooling 1" experiment. Correspondingly, in the "Warming" case,

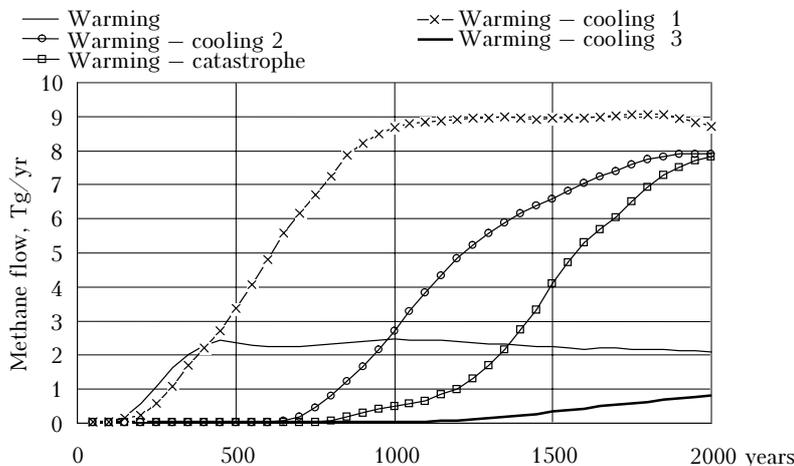


Fig. 1. Methane flow from the ocean to the atmosphere due to destabilization of the methane hydrates, Tg/yr.

Table. Comparison of parameters of subsea methane hydrates in experiments

Experiment	Duration of experiment, yr	Beginning of destabilization, yr	Maximum methane flow, Tg/yr	Time for achievement of the maximum flow, yr	Maximum methane concentration, ppb	Time for achievement of the maximum concentration, yr	Maximum bottom area occupied by decomposing hydrates, %	Time for achievement of the maximum area, yr
Warming	2000	76	2.5	438	595	908	6.2	305
Warming – Cooling 1	4000	76	9.1	1715	2055	1307	37.4	732
Warming – Cooling 2	2000	560	7.92	2000	1880	1783	18.9	1600
Warming – Cooling 3	8000	974	1	2402	290	2324	2	2014
Warming – catastrophe	5000	717	10	4817	2172	5000	41.5	4997

the source "turned on" within the area of no more than 6% of the ocean bottom and provided for the maximum mean concentration of dissolved methane equal to 595 ppb for 908 years, while in the "Warming–Cooling 1" experiment the sources "turned on" within the area larger than 37% of the ocean bottom and provided for the maximum mean concentration of dissolved methane of 2055 ppb for 1307 years.

If the methane hydrate depth H_h increased to 300 m in the "Warming–Cooling 2" experiment, then the destabilization began already 560 years later after the beginning of the experiment, the sources "turned on" within the area smaller than 19% of the bottom and provided for the maximum mean concentration of dissolved methane equal to 1 880 ppb in 1783 years. The maximum methane flow into the atmosphere achieves 7.9 Tg/yr in only 2000 years.

If, in addition to $H_h = 300$ m, the temperature tolerance dT_h was increased to 0.1° in the same "Warming–Cooling 3" experiment, then the destabilization began only after 974 years, the maximum methane flow was only 1 Tg/yr in 2400 years, the

sources turned on only within 2% of the ocean bottom area, and the maximum mean methane concentration was only 290 ppb.

In the last experiment, the warming scenario was changed: the instantaneous catastrophic increase of the ocean surface temperature by 3° and further establishment of these temperature conditions were assumed. The other parameters were the same as in the previous experiment: $H_h = 300$ m; $dT_h = 0.1^\circ$. The process of destabilization of the methane hydrates began 717 years after this, which resulted in the gradual increase of the methane flow into the atmosphere during the whole experiment. The methane flow became 8 Tg/yr in 2100 years and for the integration time up to 5000 it increased to 10 Tg/yr.

The analysis of the calculated concentrations of dissolved methane in the ocean (Figs. 2 and 3) shows that methane hydrates, located at a depth of up to 1000 m in high latitudes in the Greenland Sea near the coast of North America, near Antarctica, and in the Arctic basin, destabilize in the first turn after the beginning of warming.

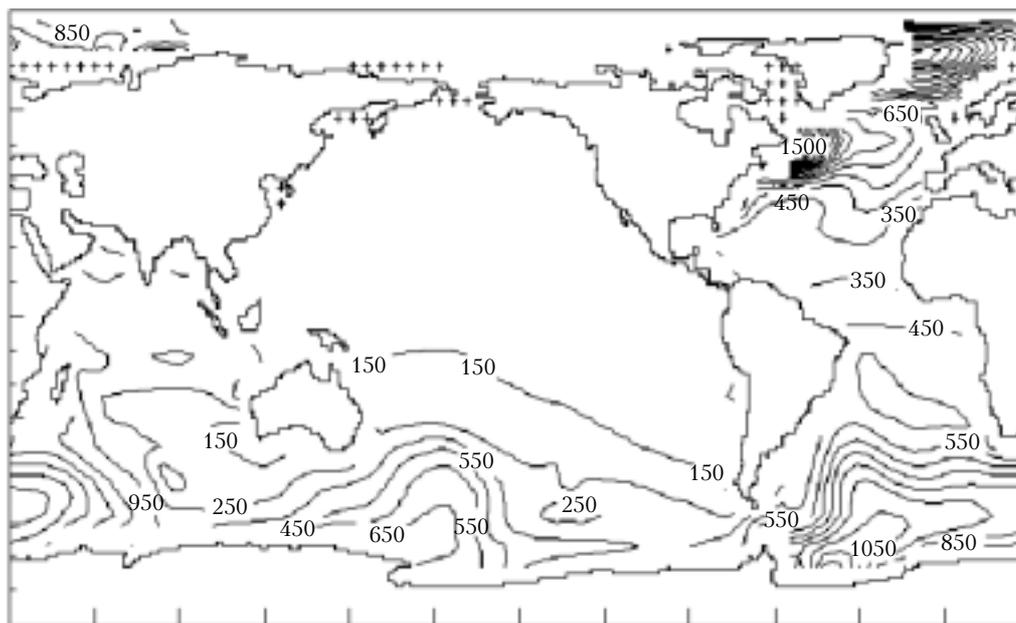


Fig. 2. Distribution of dissolved methane (ppb) in the ocean at a depth of 500 m in the "Warming" experiment in 400 years after the beginning of experiment.

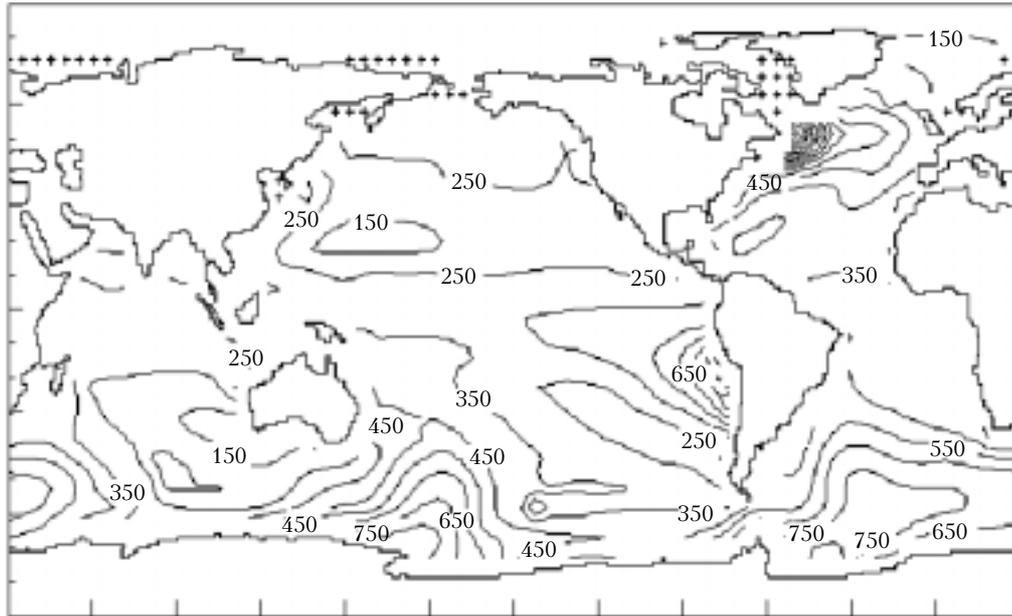


Fig. 3. Distribution of dissolved methane (ppb) in the ocean at a depth of 500 m in the "Warming–Cooling" experiment in 1000 years after the beginning of experiment.

The experiments conducted with the long integration time have shown that a relatively short secular-scale disturbance of the ocean surface temperature can lead to a long, with the period up to 5000 years, saturation of the World Ocean water with methane.

In general, the experimental results demonstrate a reasonable agreement in the methane flows and the time of beginning of destabilization with the estimates available.¹⁰

Acknowledgments

This work was supported by the SB RAS Integration Grant No. 147 "Natural and Synthetic Gas Hydrates."

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