On possible increase of the methane flux into the atmosphere from the destabilized methane-hydrates under changes of the ocean surface temperature

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The possibility of methane-hydrate decomposition due to global climate changes is investigated using the world ocean model. It is supposed that the temperature of the ocean surface rises up to 1.5° during 50 years and then decreases to 1.5°. The climatic model of the ocean is integrated for the period of 2500 years and simulates penetrations of warming from the surface deep into the ocean. A three-dimensional equation of transport of dissolved methane from sources is simultaneously solved. The methane flux into the atmosphere has reached from 4.4 to 7.3 Tg/year, depending on a warming scenario.

In the previous works [1–3], the problem of the possible consequences of destabilization of methane-hydrates which contact with the sea water under an increase of the surface ocean temperature was considered. It was supposed that methane-hydrates exist at the bottom and at the continental slope everywhere, where the pressure-temperature conditions of their stability are valid. As soon as warming from the ocean surface reaches some nearbottom waters and raises their temperature on definite value – tolerance [4], the destabilization of methane-hydrates begins. In the model, the destabilization is set as an increase in concentration of methane at a nearbottom grid point. It appears, that at such assumptions the time interval from the beginning of surface waters warming to the beginning of a possible destabilization is small, about several years because in high latitudes the warming from surface waters due to the vertical convection quickly reaches the nearbottom waters.

The data on distribution of the ocean methane-hydrates show that as a rule they lie in a sedimentary layer at subbottom depths about hundreds meters. Therefore, for the decomposition of methane-hydrates the penetration of warming not only to the bottom, but also into a sedimentary layer, which demands a greater time in comparison with the case of methane-hydrates contacting directly with water, is necessary.

The data about paleoclimate [5] show that climatic changes have a periodic character. Warming is replaced by cooling and the periods of these

fluctuations change from tens up to thousands years. As opposed to the previous results [1–3], in this work scenarios of changes of the surface water temperature both with warming and with cooling with a period of 100 years are considered.

The climatic state of the ocean was obtained from the three-dimensional World ocean model including seasonal variability, with real bottom topography, [6]. The problem was solved in the polygonal area of the World ocean on the five-degree latitude-longitude grid from 72.5°S up to 87.5°N with 24 vertical levels till the stationary state through the time about several thousands years, [7].

The equations describing the ocean climate look like:

$$R_1 u + \ell v = -\frac{1}{a\rho_0 \sin\theta}\frac{\partial P}{\partial \lambda} + \frac{\partial}{\partial z}\nu\frac{\partial u}{\partial z}, \tag{1}$$

$$-\ell u + R_1 v = -\frac{1}{a\rho 0}\frac{\partial P}{\partial \theta} + \frac{\partial}{\partial z}\nu\frac{\partial v}{\partial z}, \tag{2}$$

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

$$P = -g\rho_0\zeta + g\int_0^z \rho\,dz, \tag{4}$$

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

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

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

Boundary conditions:

$z = 0$:

$$v\frac{\partial u}{\partial z} = -\frac{\tau_\lambda}{\rho_0};\ \ v\frac{\partial v}{\partial z} = -\frac{\tau_\theta}{\rho_0};\ \ w = 0;$$

$$T = T^*(t, \lambda, \theta, 0),\ \ S = S^*(t, \lambda, \theta, 0); \tag{8}$$

$z = H$:

$$v\frac{\partial u}{\partial z} = -R_2\int_0^H u\,dz,\quad v\frac{\partial v}{\partial z} = -R_2\int_0^H v\,dz,\quad w = \frac{u}{a\sin\theta}\frac{\partial H}{\partial \lambda} + \frac{v}{a}\frac{\partial H}{\partial \theta},$$

$$\kappa\frac{\partial T}{\partial z} = 0,\quad \kappa\frac{\partial S}{\partial z} = 0. \tag{9}$$

On the lateral surface Γ:

$$\mu\frac{\partial T}{\partial n} = 0,\quad \frac{\partial S}{\partial n} = 0,\quad u_n = 0. \tag{10}$$

At the initial time moment

$$t = 0: \quad T = T^*(\lambda, \theta, z), \quad S = S^*(\lambda, \theta, z). \tag{11}$$

The following notations are used: u, v, w are velocity vector components on the coordinates λ, θ, z, where λ is longitude, θ is the addition of longitude up to $90°$, z is directed vertically downwards; t is time; P is pressure; ρ_0, ρ are the mean value and the anomaly of density; $\zeta = \xi - P_{am}/(g\rho_0)$ is the reduced level; P_{am} is atmosphere pressure; $z = \xi(\lambda, \theta)$ is the equation of ocean surface; $R_1 u$, $R_1 v$ are the parameterization of the horizontal turbulent viscosity; ν is the vertical turbulent viscosity coefficient; k, μ are the vertical and horizontal turbulent temperature and salinity diffusion coefficient; $\ell = 2\varpi \cos\theta$ is the Coriolis parameter; $\tau_\lambda(t, \lambda, \theta)$, $\tau_\theta(t, \lambda, \theta)$ are the known wind stress; $T^*(\lambda, \theta, z)$, $S^*(\lambda, \theta, z)$ are the calculated three-dimensional climatic distribution of temperature and salinity, obtained after the integration of model (1)–(7) till the stationary state; $T^*(t, \lambda, \theta, 0)$, $S^*(t, \lambda, \theta, 0)$ are the known seasonally changing surface temperature and salinity; R_2 is the coefficient of bottom friction; n is the normal to the lateral cylindrical wall Γ; $H(\lambda, \theta)$ is the bottom relief.

The ocean model is supplemented with one-dimensional model of a bottom sedimentary layer:

$$\frac{\partial T_s}{\partial t} = \frac{\partial}{\partial z} k_s \frac{\partial T_s}{\partial z}, \tag{12}$$

$$z = H: \quad T_s = T_{s,\text{climate}}, \tag{13}$$

$$z = H + d: \quad k_s \frac{\partial T_s}{\partial z} = Q_t, \tag{14}$$

$$t = 0: \quad T_s = T_{s,\text{climate}}(z). \tag{15}$$

Here $k_s = 10^{-2}$ cm/sec is the conductivity coefficient of the sedimentary layer; $Q_t = 3°/100$ m is the geothermic heat flux; $T_{s,\text{climate}}(z)$ is the solution of the stationary equation (12) with the climatic temperature from problem (1)–(10).

The transport of the dissolved methane is described by the equation

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

with the boundary conditions

$$z = 0: \ C = C^*(\lambda, \theta, 0); \quad z = H: \ \kappa \frac{\partial C}{\partial z} = 0; \quad \text{on } \Gamma: \ \mu \frac{\partial C}{\partial n} = 0. \tag{17}$$

Here C is the concentration of the dissolved methane, $C^*(\lambda, \theta, 0)$ is the known values of the ocean surface methane concentration, which zonally

change from 50 ppb in southern high-latitude areas of the ocean up to
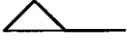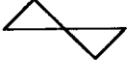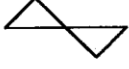100 ppb in northern high-latitude areas.

A few scenario experiments were carried out. Each experiment began
with the same initial calculated climatic fields. As well as earlier, it is sup-
posed that methane-hydrates exist in a sedimentary layer everywhere, the
pressure-temperature conditions of their stability are valid. In each exper-
iment, the time of the beginning of destabilization of methane-hydrates, a
part of the bottom containing such sources, and the diffusive methane flux
into the atmosphere are determined.

In the experiment "Warming", the seasonally changing surface ocean
temperature everywhere, except for the polar areas, covered with ice, all
year round linearly increases by 1.5° for the first 50 years of the experi-
ment, and then it decreases also linearly by 1.5° up to the initial values,
also for 50 years. Further the ocean surface temperature continues to vary
according to the seasonal climatic data. A thermal signal from the ocean
surface by the convective mixing and currents is transported to the deep
ocean up to the nearbottom layers. Further, with the help of the one-
dimensional equation of heat conductivity (16) with conditions (17) an in-
crease in temperature of a sedimentary layer is simulated and compared to
the initial climatic values. As soon as this increase exceeds some allowable
value – the tolerance, [4], which is a parameter of the scenario, for example,
Tol = 0.01°, also there will be it on determined subbottom depth, for exam-
ple, $H_h = 100$ m where methane-hydrates lie and which also is parameter
of the scenario, it is supposed, that the decomposition of methane-hydrates
begins. This decomposition supports the nearbottom concentration of the
dissolved methane equal to 5 000 ppb. Thus it is considered that the methane
released by the decomposition is completely dissolved in the sea water. This
value is further used as boundary condition at this point of the bottom for
the solution of the methane transport equation, instead of a zero methane
flux.

Thus, a source of methane "is switched on" in the model. If further dur-
ing the experiment an increment of heat in a sedimentary layer at methane-
hydrates depth becomes less than the tolerance Tol, it is supposed, that the
destabilization has stopped or in other words, a source "is switched off".
This means that again there is used a boundary condition for the equation
of methane transport with a specified value 5 000 ppb.

In the experiments "Warming–Cooling" 1–3, the seasonally changing sur-
face ocean temperature changes again in the same manner during the first
100 years, as well as in the experiment "Warming", and in the following
100 years there is a cooling. Thus, the surface temperature again decreases
with a speed 1.5° for 50 years and then again increases with the same speed
up to the initial seasonal climatic values. After that the surface ocean tem-
perature again continues to change according to the seasonal climatic data.

Comparison of the possible parameters of the subbottom methane-hydrates destabilization depending on scenarios of changing the surface ocean temperature

Parameter	Experiment			
	Warming	Warming–Cooling 1	Warming–Cooling 2	Warming–Cooling 3
The scienario of the surface ocean temperature change				
Tolerance	0.01	0.01	0.01	0.1
The subbottom depth H_h, where methane-hydrates lie, m	100 in Arctic, Antarctic and 200 everywhere	100 in Arctic, Antarctic and 200 everywhere	300 everywhere	300 everywhere
Duration of experiment, years	2000	4000	2000	8000
A time of the beginning of the methane-hydrates decomposition, years	76	76	560	974
The maximal methane flux into the atmosphere, Tg/year	2.5	9.1	7.92	1
A time of reaching the maximal flux	438	1715	2000	2402
Maximum of the average methane concentration	11.9	41.1	37.6	5.8
A time of reaching the maximum of the average concentration, years	908	1307	1783	2324
A maximal part of the bottom occupied by the decaying methane-hydrates, %	6.2	37.4	18.9	2
A time of reaching maximum decaying methane-hydrates, years	305	732	1600	2014

It appears that the model of the global ocean climate responds to the cooling of surface waters stronger and longer than to warming, [8]. So, in the experiment "Warming" the ocean climate model needs about 700 years to return to the initial state, and in the experiment "Warming–Cooling" 1 it needs much more, about 1 700 years.

Accordingly, the model of methane with a sedimentary bottom layer also responds, see the table and Figures 1, 2. The time before the beginning of the destabilization in both experiments is about 76 years with the tolerance
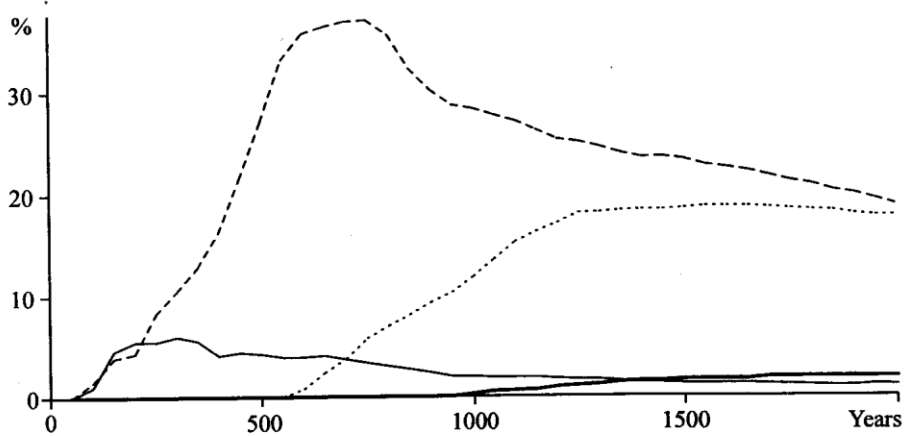
Figure 1. A part of the ocean bottom occupied by the decomposed methane-hydrates: —— Warming; − − Warming–Cooling 1; · · · Warming–Cooling 2; and —— Warming–Cooling 3
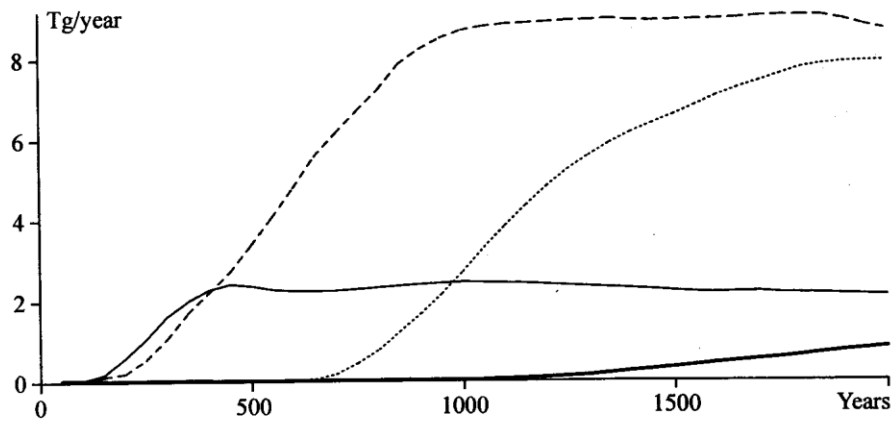

Figure 2. Methane flux into the atmosphere from decomposed methane-hydrates: —— Warming; − − Warming–Cooling 1; · · · Warming–Cooling 2; and —— Warming–Cooling 3

Tol $= 0.01°$ at the depth of methane-hydrates $H_h = 100$ m in the Southern and the Arctic oceans and $H_h = 200$ m in other oceans. The methane flux into the atmosphere reaches its maximal value 2.5 Tg/year after 438 years from the beginning of the experiment "Warming", and values 9.1 Tg/year after 1715 years from the beginning of the experiment "Warming–Cooling" 1. Accordingly, in the case of "Warming" the sources "are switched on" on the area of no more than 6% of the ocean bottom and provide a maximum of average concentration of the dissolved methane 595 ppb after 908 years whereas the sources are "switched on" in "Warming–Cooling" 1 on the area exceeding 37% of the whole bottom of the ocean and provide a maximum of

average concentration of the dissolved methane 2 055 ppb after 1 307 years.

If the depth of methane-hydrates H_h is increased up to 300 m in the experiment "Warming–Cooling" 2 then the destabilization begins after 560 years from the beginning of the experiment. The sources "are switched on" on the area less than 19maximum of average concentration of the dissolved methane 1880 ppb after 1783 years. The maximal methane flux into the atmosphere reaches 7.9 Tg/year only after 2 000 years.

If, in addition to $H_h = 300$ m, in the experiment "Warming–Cooling" 3 the tolerance Tol is increased up to 0.1°, then the destabilization will begin only after 974 years, the maximal methane flux will be only 1 Tg/year after 2 400 years, the sources "are switched on" only on 2% of the ocean bottom, and the maximal average concentration of methane will be only 290 ppb.

Analysis of the calculated dissolved methane concentration in the World ocean shows that first of all after the beginning of warming there are destabilized methane-hydrates located at depths up to 1 000 m in high latitudes in the Greenland sea and near to the western coast of South America and at the coast of Antarctica.

The carried out experiments with a long time of integration have shown, that rather a short-term perturbation of temperature of the ocean surface of century scale for a long period of time, say about 5 000 years, can lead to the saturation of waters of the World ocean by the dissolved methane.

The results of the experiments as a whole give a reasonable coincidence of the methane flux into the atmosphere and the time of the beginning of destabilization with available estimations [9].

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