Evidence and causes of strong ocean heating during glacial periods.

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Abstract

We investigated data from all initial reports of deep-sea drilling projects and have noticed that on the surface of the ocean bottom geothermal heat flux is approximately 40 mW/m^2 higher than hundreds meters below. This infers that during the last glacial period interior ocean temperature was warmer than it is today.

What phenomena can cause interior waters warming? The densest water in the world's oceans exists in the Red Sea. Under current conditions the Red Sea is slowly pumping salt down into the ocean interior and freshening the surface. Presented here model of thermohaline circulation is showing that this decrease density of water in the high latitude seas and in 1600 years northern downwelling ceases. This indicates the beginning of a new glaciation. The ocean becomes strongly stratified and only warm-dense water from the Red Sea is filling it. Additional heating to the ocean is given by the geothermal heat flux and organic rain decay. According to our calculations in 112000 years, temperature of the water in the ocean interior reaches 23.4^0C and becomes less dense than the water in the high latitude seas. This causes strong convection to occur. Water temperature in the Northern Atlantic increases by 18^0 C. Rapid glacier degradation occurs as the ocean quickly releases accumulated heat and in 14600 years ocean interior drops to 1.6^0C. After this period a new glaciation cycle starts. In our model energy input from space and earth interior is set constant but we still received asymmetrical glacial cycles.

1.1. Introduction

Numerous hypotheses have addressed glacial-interglacial climatic dynamics, but none of them explain the sharp 25^{0}C temperature increase in Greenland in the last deglaciation [Cuffey et al., 1995; Dahl-Jensen et al., 1998]. These robust data were obtained through analyzing the temperature profile in the Greenland ice sheet where cold from the last glaciation is preserved in the depth of the glacial sheet [Cuffey et al., 1995; Dahl-Jensen et al., 1998]. Approximately the same temperature rise was shown by data about the extent of vast Late Pleistocene ice wedges on the plains of western Europe [Washburn, 1979]. Such large ice wedges occur only in extremely cold environments where the average annual temperature is below -12^{0}C [Washburn, 1979]. Currently mean annual air temperature there is around +10^o C.

Planet orbit oscillations, changes in global albedo and increases in greenhouse gas concentrations are all relatively slow processes. However, the Greenland data indicate that climate changed instantly. In Greenland, ice cores reveal that until ~14650 years ago, there were no substantial fluctuations in climate [Severinghaus and Brook, 1999]. The annual accumulation of ~100 mm of snow contained relatively stable isotopic content. This signal changed sharply along with the extreme temperature perturbation. The thickness of annual ice layers doubled stepwise over 3 years as well [Severinghaus and Brook, 1999]. Then, during Younger Dryas, the climate changed back to glacial conditions, but ~11700 years ago a new sharp shift in climate occurred [Severinghaus and Brook, 1999]. It can be supposed that because of glacier retreat, change in albedo and increased emission of greenhouse gases, the temperature in northern latitudes should grow with the course of deglaciation. But data of the temperature profile in the Greenland ice sheet clearly show that the air temperature on the surface of the ice sheet has, in contrast, declined by 2.5^{0}C [Dahl-Jensen et al., 1998]. Therefore, this 25 degrees heating happened stepwise; and the highest air temperature was at that moment.

Ocean surface circulation can change in one year, but that would only cause redistributions of solar energy over the ocean surface. In order to increase the temperature in the northern Atlantic and surrounding lands by 25^0C, in some other big territories the temperature should drop respectively. 25^0C is a huge change, analogous to the difference in average annual air temperatures between the Arctic and the tropics. Such an increase is impossible to explain in the frames of existing knowledge about earth's climate system. For that, an additional energy source is required. There are many natural phenomena which display sharp asymmetric cycles similar to the glacial-interglacial cycles. Most frequently they are based on slow energy accumulation and consequent quick release [Chuprynin, 1985]. Such an accumulator could be the world ocean.

We suggest that during glaciations the ocean accumulated energy: interior ocean water heated up to ~20-30^0C and during deglaciation this energy is released. Here we consider reasons and evidences for the phenomenon.

1.2. Warm ocean hypothesis

The heaviest water is on the bottom of the ocean. Currently the ocean is filled with cold and therefore heavy (1027.8 kg/m^3) water from high-latitude seas. The salinity of the water is 34.7‰ [Hay, 1983]. But that is not a single and moreover rather rare scenario in the history of the ocean [Hay, 1983]. In the past, in warmer climates, the ocean was filled with heavy but warm water from seas with high evaporation [Hay, 1983]. Today there are two such seas (plus the Persian Gulf). First, the Mediterranean Sea, because of the difference in evaporation and precipitation where every year on average 850 mm of water is lost. Through Gibraltar straight 42320 km^3/yr of ocean water enters the Mediterranean while 40800 km^3/yr bottom flow leaves this sea with 38.4-38.7‰ salinity, and a temperature of 12.5^0C [Hay, 1983]. Annual evaporation from the Red Sea is 3500mm of water.

From it, 9400-12600 km^3/yr of water with a salinity of 41.5‰ and temperature of 22.5^0C flows down to the abyssal depths of the Indian ocean [Hay, 1983]. The densest water of the world ocean (1028-1029 g/m^3) flows out of these seas [Hay, 1983]. Both the Red and the Mediterranean seas are very deep and connect to the ocean by shallow straits. They can be considered as “small oceans”, and they can be used as examples of warm ocean interior waters [Hay, 1983]. They are filled with warm water throughout their whole depth. On the bottom of these seas, the temperature is 12.6-13.4^0C and 22.5^0C [Hay, 1983]. Today the total flow from these salty seas into the ocean interior is only several per cent of the flow from high-latitude cold seas [Hay, 1983]. Therefore their effect on ocean interior temperatures is minor; because of their influence ocean bottom temperatures are not 0^0C but $+2^0$C [Hay, 1983]. If one imagines these seas as water pumps, then these pumps have low power (debit). But these haline pumps give stronger pressure (density of water), therefore they are stronger than thermo pumps of the high-latitude seas. Their debit is small compared with a thermal pump, but the stream from the Red sea would fill the whole of the world ocean volume in 100,000 years, i.e. for one glacial period. Flow from the Mediterranean Sea would fill the world ocean 4 times faster.

One very important phenomenon is connected with the function of such haline pumps: they reduce the salinity of the ocean surface and compete both with thermal pumps and between themselves. They pump salt down to the ocean bottom and fresh water evaporated from these seas comes back to the ocean surface. They pump down the ocean bottom water with high salinity and instead somewhere in the world ocean appears a compensating flow from the depth to the surface with 34.7‰ salinity. While thermal pumps strongly mix the ocean, this process is barely noticed, but if thermo pumps weakened an irreversible freshening of surface waters would begin. One possible trigger for this can be, for example, the melting of glaciers and short-term freshening of the ocean surface [Hay, 1983]. The density of this water would reduce and the downflow of cold water would stop [Hay, 1983]; the haline pumps would continue to work thereafter; haline pumps would quickly make the surface water even fresher. Water in cold seas becomes lighter and would not yet be able to submerge. The ocean would become strongly stratified; wind and tides would not be able to mix this water with heavy bottom waters. As a result, haline pumps would pump warm and salty water onto the ocean floor without interference.

Another possible reason for ocean interior waters to warm is geothermal heat flux. Energy of $50mW/m^2$ is sufficient to warm an entire ocean water column (4km) during one glacial cycle (100,000 years) by 10^0C [Adkins et al., 2005]. Another mechanism is biological heating. About 11 Pg (1 Pg=10^{15} g) of organic matter (representing about one-quarter of the ocean bio productivity) descends annually to the ocean interior and the bottom [Chapin et al., 2002]. Only 2% of the value is buried in the bottom sediments [Chapin et al., 2002]. Taking into account the ocean area of 0.361 x $10^{15}m^2$, on average, about 30 g of carbon (or ~60 g of organic matter) is oxidized on each square metre of the ocean floor annually. Heat flow of 40 mW/m^2 is computed using an average 5 kcal/g (21 kJ/g) energy content of 'sea food'.

2.1. Proof of ocean heating in the glaciations

The conclusion that thermal convection in the glacial ocean (ocean ventilation) was significantly reduced is based on a big difference between the 14C age of planktonic and benthic foraminifera [Sarnthein, 2011]. The same conclusion was based on the tenfold increase in productivity in the Southern Ocean during the last deglaciation [Anderson et al., 2009]. Because of increased circulation, the nutrition supply from the interior to the ocean surface has sharply increased [Anderson et al., 2009].

The most exact data of surface ocean salinity in the last glaciation are obtained through analysing plankton fauna in deep boreholes of the north-western Atlantic and numerous samples obtained for recent complexes of these species from different basins in northern Atlantic and the Arctic [de Vernal and Hillaire Marcel, 2000]. Before the peak of glaciation, huge masses of fresh ice were accumulating on land and on sea surfaces, and water salinity should have been expected to rise at that time. However, on the north-west Atlantic both before and during the last glacial maximum (LGM) it was 30-32‰ [de Vernal and Hillaire Marcel, 2000]. Same analyses can't be conducted for other regions, since for comparison fauna from seas with low salinity are needed, while such seas exist only in the north. The northern Atlantic is an open ocean. Water on the surface of the world ocean is mixed quickly. From this it follows that on the entire ocean surface in LGM salinity was ~3-5‰ lower than today. We have received the same value in our model (see below).

Even cold water with such salinity is very light. No tides or winds can submerge such water to the ocean floor. Geothermal energy and biogenic heating are real phenomena. If the entire ocean was strongly stratified, then there would have been no vertical mixing. Diffusion heat losses are minor, therefore these two sources must have warmed the ocean interior.

If the ocean surface was freshening, then on the ocean interior salinity must have risen. And this anomaly must have persisted in the depth of the bottom sediments. And in truth, in various regions of the ocean high-salinity pore water was detected. This allowed water salinity to be estimated on the ocean bottoms in the LGM. It appeared to be anomalously high – ~36.5‰ [Adkins et al., 2002]. Such salinity can't be explained solely by glacier growth [Adkins et al., 2002].

If the temperature and salinity of water on the ocean bottom in the past often and strongly changed (during all glacial cycles), then benthic organisms should be resistant to these changes. This is supported by the present distribution of foraminifera on the ocean floor. This is indeed not connected with temperature and salinity and depends on food supply [Van der Zwaan et al., 1999] The same species can now be found both in the warm bottom of the Red Sea and in the cold bottom of the Norwegian Sea [Badawi et al., 2005].

Our hypothesis supposes that in the LGM the ocean floor was filled with warm water from the Red Sea. This means that the temperature and $\delta^{18}O$ of water on the bottoms of all oceans should be close to those of the Red Sea. That means that benthic foraminifera $\delta^{18}O$ should be close as well. Analyses of δ18O of benthic foraminifera showed that on the warm bottom of the Red Sea in the

LGM δ18O was +(4.7-5.5)‰, the same as in the bottom of all other oceans and the same as δ18O in plankton foraminifers of the northern part of the Red Sea where deep water forms [Luz and Reiss 1983; Hemleben et al., 1996; Geiselhart, 1998; Badawi et al., 2005]. This confirms that water from the Red Sea had spread along the bottom of all oceans.

And now let's discuss the same fact but assuming that the temperature on the ocean floor in LGM was close to freezing everywhere. Then it appears that the δ^{18}O paleothermometer is not always correct. For the LGM it shows the same signal both for the cold ocean and for the warm Red Sea. It is possible that biological fractionation of isotopes on the ocean floor at high temperature and salinity, poor food and oxygen source and high carbonate ion concentration differed from the present. Carbonate composition of benthic foraminifers' shells in the LGM likely also differed from the present. Possibly, for the LGM ocean floor condition, it should be used not calcite scale of δ^{18}O paleo thermometer, but aragonite scale.

Beside that, the δ18O paleothermometer doesn't account for water fractionation in the Red Sea. Isotopically light water evaporates there and returns to the ocean surface with precipitation, while isotopically he avy water submerges to the ocean floor. Today this process is hard to detect, but in the LGM, when ocean mixing was low, this was an important factor. The same with salinity: because of fractionation in the Red Sea, δ18O on the ocean floor in the LGM was higher than on average in the world ocean, and on the surface it was lower. (Below we will provide a new estimate of water temperature on the shelf of New Zealand in the LGM. Based on this, the ocean surface was cooling as strongly as terrestrial regions did). Below we will return to the δ^{18}O paleothermometer.

Another paleothermometer is based on the ratio between Mg and Ca in shells. Dependence of this thermometer on temperature is a matter of debate. This paleothermometer often shows either high temperatures in the ocean interior in the LGM [Bryan and Marchitto, 2008] or a complete disconnection with temperature [Bryan and Marchitto, 2008; Yu and Elderfield, 2008]. We are not going to discuss this method in more detail here. We have a more reliable method to measure temperature on the ocean bottom in the LGM. This can be done by directly measuring temperature in the bottom sediments in the deep boreholes.

2.2. Deep borehole paleothermometer

If for an extended period of time the temperature on the bottom was high, then bottom sediments must have warmed as well. When cold water penetrates to the bottom, sediments will start to cool. The top 5 meters will cool quickly – within a year, sediment layers 10 times thicker (50m) will cool 100 times longer, while 500 meters will cool for 10,000 years [Yershov, 2004].

The cold have persisted at a depth of the Greenland ice sheet from the last glaciation. This allowed the temperature on the surface to be estimated accurately in the LGM [Dahl-Jensen, 1998]. If in the LGM the ocean bottom was 25-30^{0}C warmer than in the bottom sediments in the depth of hundreds of meters this heat must have persisted. Detecting this anomaly by looking at the temperature profile and estimating the temperature dynamic in the past is easier for the ocean borehole than glaciers. In glaciers, active snow accumulation takes place, glaciers flow and all that should be accounted for in the calculations. On the ocean floor, sediments usually accumulate slowly, usually parts of a millimeter per year. Sediments stay intact.

Now nobody is assuming that the ocean bottom in the LGM could be warm. But 40 years ago when *Glomar Challenger* was just starting to drill the boreholes, this was actively discussed [Erickson et al., 1975]. By that time over 3,000 measurements of heat flux through the ocean floor had been made, and it was necessary to find out whether all this flux is coming from deep crust interior or this is mostly the release of heat which was accumulating during the last ocean warming [Erickson et al., 1975]. In the deep boreholes it was easy to measure. If there was no change in water temperature in the past then heat flux in all boreholes should not change with depth and should be equal to geothermal heat flux [Erickson et al., 1975] (heat flux is calculated as a multiplication of the heat-conductivity coefficient by temperature gradient [Erickson et al., 1975]); and if in the last glaciation the ocean was warmer, then heat flux should be maximum at the bottom surface and decline with depth in bottom sediments [Erickson et al., 1975]. Only at the big depth the flux should be equal to the geothermal heat flux [Erickson et al., 1975]. At the bottom surface, the flux of cooling sediments should be added to the geothermal flux [Erickson et al., 1975].

The Atlantic, Indian and Pacific oceans connect by deep and wide straits. Therefore, water in these oceans had and now has similar temperature [Hay, 1983]. In such a case there is no need for many deep boreholes to test this hypothesis. Even one accurately measured heat flux profile is enough (for example, as for temperature reconstruction in Greenland). If the temperature of the ocean in the past didn't change substantially, then all boreholes should show stable with depth heat flux [Erickson et al., 1975]. And if in the LGM the ocean was 25-30^{0}C warmer than in all boreholes, heat flux at depths should be by several tens of mW/m2 lower than at the ocean bottom surface [Erickson et al., 1975].

In the frames of the Deep Sea Drilling Project (DSDP), *Glomar Challenger* has done 96 legs; the heat flux profile was measured in numerous deep boreholes in the different parts of the world ocean, and as noted in the final reports [Erickson et al., 1975; Hyndman et al., 1987], heat flux was constant with depth. This means that the temperature of the ocean interior didn't vary substantially in the past, and the observed heat flux at the ocean bottom is a geothermal heat flux [Erickson et al., 1975; Hyndman et al., 1987].

As far as we know after publication of these authoritative papers, possible ocean interior warming in the LGM was not discussed further in the literature. Stability of temperature in the bottom in the Pleistocene has become a well-proven scientific fact. Models of ocean circulation, climate, global carbon cycle and models of convection in the mantle were made based on this fact. However, we critically read these final reports [Erickson et al., 1975; Hyndman et al., 1987] and in addition carefully investigated numerous initial reports of the DSDP, which were used in the preparation of final reports, and have found striking discrepancies.

Geothermal measurements made from the *Glomar Challenger* over the first 5 years of the DSDP to leg 26 have been reviewed by Erickson et al. [1975]. The main goal of the review was to find out whether these data correspond with depth geothermal heat flux, if it wasn't biased with Pleistocene-Holocene bottom water temperature dynamic [Erickson et al., 1975]. We will briefly repeat the peculiarities of these investigations.

2.3. Feature of heat flux measurements in the deep boreholes

During the drilling of boreholes, circulation of cold water (drilling fluid) in the borehole strongly changes the temperature regime of the sediments. It takes several weeks in order for the water to be in equilibrium with surrounding sediments. Ocean ships cannot wait so long [Hyndman et al., 1987]. Therefore temperature measurements are conducted with a down-hole temperature recorder, which is put down the borehole, and using high pressure a thin probe with a temperature sensor is put into the thermally undisturbed sediments [Hyndman et al., 1987]. The sensor usually measures temperature every few seconds. On the data file based on temperature registration can be seen as the recorder is going down the cold hole's bottom, as the temperature of the probe quickly rises after the probe has penetrated the undisturbed sediments, as the temperature of the probe after that slowly reaches equilibrium with the surrounding sediments (ideally reaches plateau), and then as the probe is removed, and appears in cold water again. Frequently it can be seen as the probe is getting heated penetrating into the dense sediments with friction, and then becomes balanced with the surrounding sediments and cools down. If thermo equilibrium isn't reached over the course of measurements (tens of minutes) then measurement is extrapolated up until to equilibrium values [Erickson et al., 1975; Hyndman et al., 1987].

Having temperature measurements in different levels it is possible to calculate the average temperature gradient between these levels (temperature difference divided by depth difference) [Hyndman et al., 1987]. The average heat flux for each depth interval is calculated by multiplying the temperature gradient in this interval and the coefficient of heat conductivity [Hyndman et al., 1987]. If there are several measurements of heat conductivity in this interval then average value is taken. Heat conductivity in turn is measured in cores taken on board, from the borehole [Hyndman et al., 1987].

In the final report, Erickson et al. [1975] have provided temperature profiles of bottom sediments, calculated for different hypothetical scenarios of Pleistocene-Holocene ocean interior temperature dynamic. We have repeated these calculations and our profiles in general corresponded with profiles presented in Erickson et al. In Figure 1A we show the temperature dynamic of bottom sediments for the following scenario: at the beginning of each glacial cycle the temperature of the water on the ocean bottom is equal to $+2^{o}$ C, and then in the course of 100,000 years it gradually (linearly) increases to $+32^{o}$ C. The profile of the bottom temperatures at that time (period of maximum heating) is shown with red lines. After that, the temperature of the water abruptly decreases back to $+2^{o}$ C, and sediments begin to release heat back to the water. Green lines indicate the temperature profile after 11,470 years of ocean cooling and blue after 14,450 years. On the right side of Figure 1A corresponding heat flux profiles for these time intervals are shown. Calculations are made for 3 different values of geothermal heat flux – 0 mW/m^2, 14 mW/m^2 and 42 mW/m^2; constant with depth heat conductivity coefficient – 1.4 mW/m ^{0}C, which is the most typical value (see Figure 2), and a heat capacity of 3.35 J/cm^3 ^{0}C, which for sediments varies in a narrow range [Yershov et al., 2004].

In Figure 1A we see bottom sediments cooling after cooling of the ocean. Most heat is released from the near surface sediments. They have cooled by 30^{o} C quickly. Sediments at a depth of 500 meters have cooled in 14,450 years by 10^{o} C. Sediments at a 1000-metre depth have cooled only by 2^{o} C. These sediments began to cool only after 12,000 years of cold ocean. In Figure 1A we see that for all 3 variants of geothermal heat flux, heat flux near the surface exceeds geothermal heat flux by 35 and 44 mW/m2, and temperature profiles in cooling sediments have salient shapes. In Figure 1C temperature profiles for the same conditions but for the stable water temperature at the bottom (+2^{o} C) are presented for comparison. In this case heat flux is constant with depth and temperature profiles are straight lines. If to accept heat conductivity vary with depth then heat flux will still stay constant, but temperature profiles will become curved [Hyndman et al., 1987].

In many regions, gas clathrates are preserved in the ocean floor sediments [Miles, 1995]. At a depth of 2 km and temperature of ~17^{o} C, gas clathrates either melt or freeze (condition of phase shift). At a depth of 6 km phase shift occurs at a temperature of ~25^{o} C [Miles, 1995]. In Erickson et al. [1975] this effect has not been taken into account. If the temperature of bottom sediments has been and is stable, then it should be impossible to detect gas clathrates or methane in the sediments, looking at the thermal flux profile [Yershov, 2004]. It will not change with depth. However, if temperatures had increased by over 25°C during the glaciation, then most of the gas clathrates would have melted, and frozen back after ocean cooling. If the latter is true, then we should be able to locate a boundary phase change (“permafrost floor”) at a depth of a few hundred meters and this boundary should be slowly moving down (several millimeters per year). This is a full analogue of real permafrost dynamic [Yershov, 2004]. During gas clathrates crystallization approximately 500 J/g is released [Yershov, 2004]. This is more than at water freezing. Therefore, at freezing or thawing of gas clathrates releases or consumes lots of heat at its roof or foot resulting in a sharp shift in the heat flow profile and a bend in the temperature profile. In Figure 1B we show profiles with the same conditions as in Figure 1A, but include a third of the sediment volume as gas clathrates or oversaturated with methane water. For these scenarios we have accepted 20-21°C temperatures for that of gas clathrates melting (that is 3 km depth of ocean). To ease calculations, we have replaced the conditions of phase change with the condition that in the range of temperatures 20-21°C the heat capacity of sediments is not equal to 3.35 J/cm3°C but 40 times higher. We see that the presence of gas clathrates increases the difference between heat flux near the surface and in the deeper horizons. In deep sediments in this case negative heat flux may even occur (heat flux not towards the ocean bottom but to the mantle) (Figure 1B).

The same as terrestrial permafrost, "permafrost" in the bottom sediments can be two-layered. Let's assume that the gas clathrates layer is situated at depths from 0 to 600 meters below the ocean floor. At warming, 400 meters of it melts from the top and 20 meters from the bottom (because of the geothermal heat flux). During the next cooling, near surface sediments begin to "freeze" anew. Under this new "frozen" layer will be situated a layer of "relic permafrost".

If strong ocean heating occurred during glacial periods, then bottom sediment temperature profiles throughout the ocean should mirror those in Figure 1A,B. If temperature changes were less than 30°C, then related heat fluxes were also correspondingly less [Erickson et al., 1975]. If the gas clathrates content in sediments was less, then a leap in heat flux would be observed deeper. If bottom cooling happened slowly – over several thousand years – then heat flux on the surface will increase by 5-15 mW/m^2 as compared with Figure 1A [Erickson et al., 1975]. If the ocean temperature did not change, then the profiles must be as in Figure 1C in spite of the presence or absence of gas clathrates.

Increased flux of heat near the surface is connected with the release of previously accumulated heat to ocean waters. Theoretically, the same profiles as in Figure 1A can be obtained if inside these sediments other energy sources are present with a total power of tens of mW/m^2 (for example, radioactive or biological decay). However, in reality this explanation is unlikely. Such a flux can be created by tens of kilometers-thick granite strata [Pollak et al., 1993]. In the layer of carbonate silts with only a few hundred meters thickness, there can't be enough radionuclide's to maintain such a heat flux. Above we showed that the total average heat release from aerobic organic decomposition on the ocean floor is 40 mW/m2. But most of this heat is released at the very top of the bottom sediments. Only an extremely minor part of the energy penetrates to the deep sediments, and its effect can be ignored.

Not all boreholes can be used for paleotemperature reconstructions. For example, near the rift zones heat flux is very strong – hundreds of mW/m^2. Precision of heat flux measurement there is comparable with the value of possible change in heat flux with depth, besides that heat transport there occurs not only by conductivity but also with convection [Hyndman et al., 1987; Yershov, 2004]. Where pore water is ascending to the surface, the temperature profile is salient; where cold water is descending down the sediments, it is arched [Hyndman et al., 1987; Yershov, 2004]. Shallow boreholes give little information. In Figure 1A,B it can be seen that a bend in the temperature profile becomes noticeable at depths deeper than 400 meters. Approximately at the same depths gas clathrates boundary should be expected. Precision of heat conductivity coefficient estimates usually lie within the 5-15% range [Hyndman et al., 1987]. This coefficient often strongly varies with depth (see Figure 2). Therefore errors in heat flux profile calculations in shallow boreholes exceed the possible effect from the temperature dynamic.

A series of international programs have measured deep sediment temperatures from boreholes drilled throughout the world's ocean basins. These include the Deep Sea Drilling Project (DSDP), legs 1-96, the Ocean Drilling Project (ODP), legs 100-210, and the Integrated Ocean Drilling Project (IODP), legs 301-340. For each leg, the results of temperature, heat conductivity and heat flow measurements in boreholes and geological settings were published in the corresponding volumes of "Initial Reports", in the chapter dedicated to that borehole (site) and/or in a special section dedicated to heat fluxes. We revised the data of all depth boreholes. All referenced data of holes below are taken from corresponding initial reports.

2.4. Results of heat flux measurements in the deep boreholes

Erickson et al. [1975] reviewed data from 12 first boreholes (DSDP) from which 2 or more temperature measurements were collected (plus bottom temperature data; inflection of curve is visible at least by 3 points). The authors of the review took temperature data, conductivity, gradients and heat flow data from the initial reports and combined them onto a sheet. We display all these data in Figure 2A (sites 194-254). All black dots in that figure are taken from review tables (we changed nothing in these). We excluded from the analyses only the most shallow borehole (site 209) (52 meters) and three shallow boreholes drilled in the rift zone of the Red Sea. They show very high heat fluxes (115-300 mW/m^{2}yr [Erickson et al., 1975]); these data are beyond the figure data range). At such high and variable fluxes it is impossible to detect bottom temperature dynamics within these shallow boreholes. Locations of investigated boreholes we present in figure 3.

Erickson et al. used part of the data from the sheet for the final heat flux graphs. We present (repeat) the original data from Erickson's heat flux graph with orange outline (Figure 2A). The vertical size of the outline represents the extent of the averaging interval and the horizontal size presents probable error estimates made by the authors of the review [Erickson et al., 1975].

We initially note that in this graph, boreholes 184, 185 and the lower part of borehole 206 weren't included. These 3 holes are the very first boreholes, on which the heat flux profile was based. They are deep (300-600 m) and all 3 indicate a several times decline in heat flux with depth. These data were excluded from consideration by Erickson. And in the final graph based on which all conclusions were made only shallow boreholes were included. Erickson has compared those profiles with theoretical profiles (see Fig. 1A), obtained for various dynamics of bottom ocean water temperature [Erickson et al., 1975].

Based on the data, Erickson et al. [1975] concluded: "there is no consistent indication of a significant vertical increase or decrease in heat flux, such as might be caused by long-term changes in bottom water temperature"; (p.2515) "the heat flux usually remains constant within the estimated probable error of the individual heat flow determination with depth" (p. 2527). This conclusion corresponds with the final graph. On it (everything in orange in Figure 2A) we see that three boreholes have indicated decreasing heat flux with depth (206, 253, 254), and two have indicated increasing heat flux with depth (214, 217). And one borehole has indicated stable heat flux with depth (242). So the conclusion of Erickson may seem to correspond with the data. But firstly, the lowest point for borehole 271 in the final figure of the review is a mistake (The orange dot in Figure

2A is a mistake). The actual value is a black dot located to the left. The value is shown both in the initial report and in the review table [Erickson et al., 1975]. Therefore the heat flux in this borehole in reality declines with depth and not vice versa. This drastically change the statistics – now 4 boreholes indicate decreasing heat flux with depth with only 1 (214) indicating increase, but the authors of the review noted the anomalously high heat-conductivity coefficient in the lower part of the borehole, 1.3 W/m^{o}C (see Figure 2A). The authors suggest that: "These high conductivity values may have been affected by convection of interstitial water during the conductivity measurement (sediments from the lower interval were noted for their unusually high water content) rather than by an actual increase in the *in situ* conductivity" (p. 2523). This site should be either excluded from the analyses or its heat-conductivity coefficient should be corrected. This coefficient can be roughly estimated [Schloessin and Drovak, 1977]. Heat conductivity of sediments as a first approximation is combined from water heat conductivity (0.6 W/m°C) and mineral heat conductivity (1-3 W/m°C) [Nobes et al., 1991; Yershov, 2004]. Therefore, the higher the water content the lower the conductivity. If porosity (watering) of sediments increased with depth in the borehole (it sounds in the initial report "like soup"), then most likely heat conductivity would decrease with depth [Nobes et al., 1991]. Temperature gradients were constant by depth in this borehole, therefore heat flux likely decreased with depth.

Among all of the data examined in the Erickson et al. [1975], only one site (site 242) had no changes in heat flux with depth (see Figure 2A). Measurement on the site at a depth of 141 m got the highest reliability grade (Figure 2A). But do not trust this. The authors of the review noted that: "our error estimates are very subjective" (p.2519). Here is what we read about the measurement in the initial report: "On the first run (site 242), the latching device was attached to the extender with set screws to prevent the DHI (downhole instrument – Z., Z) from being pushed up into the inner core barrel. The latch did not return with the DHI but was recovered as coring continued." And further: "The record for the first run (Figure 1) indicates a relatively constant temperature of 6.25^{0}C after the bottom of the hole was reached. There is no indication of heating due to friction at the pull-out. These observations, coupled with the fact that the release latch on the probe came off, suggest that the probe was pushed up into the inner barrel before it could penetrate the undisturbed sediment. However, heat-flow calculations indicate that the temperature measured may be very near the ambient temperature at that depth" (p. 349). From the description it follows that the temperature of the water in the hole's bottom was measured at a depth of 141 m, but not the temperature of the undisturbed sediment. Nevertheless, the measurement wasn't excluded from the analyses and was even given the lowest error grade. This was done because at value 6.25^{o}C heat flow calculations indicate stable by depth flux. But it is impossible that the waters from an only just drilled borehole are near ambient sediment temperatures as thermo equilibrium requires several weeks at least [Hyndman et al., 1987]. This suggests that the temperature at this site at 141 m is significantly (by several degrees) higher than 6.25°C. The dot on the left graph of Figure 2A should be shifted to the right (blue line) and then the temperature gradient in the upper part of the borehole will increase, decreasing in the lower part (blue arrows in Figure 2A), with a resultant decrease in heat flux with depth (as represented by blue arrows in Figure 2A). The fact that the temperature measurement at this point is incorrect (lowered) could be guessed by noticing the anomalously low resulting heat flux in this borehole – 29 mW/m^2 [Erickson et al., 1975]. Figures 2A, B, C shows that the heat flux on the floor surface everywhere exceeds 40 mW/m^2. By increasing the temperature at the 141-metre depth by several degrees we increase the heat flux on the bottom to the typical observed values.

Finally, we see that out of 6 boreholes presented in Erickson et al., 5 indicate approximately 5mW/m^2 per hundred meters decline in heat flux. In the sixth borehole (site 214) the heat flux is likely declining as well. All these boreholes are shallow and for each of them the error when estimating the heat flux gradient is comparable to the values of errors obtained during measurement. But 6 boreholes is already the reliable number for statistics. All these boreholes are drilled in different parts of the world ocean and already, based solely on these boreholes, the conclusion can be drawn that the heat flux in the bottom sediments most likely declines with depth.

Now let's investigate data from boreholes excluded from the Erickson et al. analyses.

Two deep measurements on site 206 from the Erikson et al. review indicate strong decreases in heat flux with depth (Figure 2A) and were excluded from the final heat flux graphs only for this reason [Erickson et al., 1975]. The authors did not believe that this is possible [Erickson et al., 1975]. The quality of these measurements is far from excellent; however, as suggested by Von Herzen in the initial report, the fact that two measurements from the same core show low heat flux increases their reliability. This author analysed all the possible reasons that could lead to a downward decrease of heat flux: internal radiogenic or biogenic heating in the upper intervals and upward percolation of interstitial waters could not explain the natural patterns observed. Large changes in bottom water temperatures could result in a downward decrease in heat flux, but Von Herzen could not find an explanation for how such a big ocean water temperature change could occur.

Measurements from sites 184 and 185 also indicate strong decreases in heat flux with depth and were therefore considered unreliable, and thus fully excluded from further analyses [Erickson et al., 1975]. However, firstly, these boreholes are deep, and when estimating errors in gradients, the sum of errors in temperature measurements is divided by the difference in depth. Therefore, the greater the distance between the two measurements (the deeper the borehole) the more reliably the temperature gradient is likely to be calculated. Secondly, both boreholes are drilled in one place on the Umnak Plateau in the Bering Sea under similar conditions. Measured heat conductivity and geothermal heat flux at depth are similar in both boreholes [Erickson et al., 1975], increasing the relative reliability of the measurements. Thirdly, the patterns in the data from all four temperature measurements appear to be of good quality: after the penetration of the probe into the sediments the temperature quickly increased before reaching a plateau in 3-4 minutes. These boreholes show strong heat flux increases at 200 m deep. We see such a profile in Figure 1B. An oversaturation of pore water with methane was recorded whilst drilling both boreholes. The bend in the temperature profile in this “combined” borehole is located on the cross point between the temperature profile and the profile of the temperature of gas clathrates freezing (Figure 2A). Most likely today, this site contains gas clathrates – above this bend sits the gas clathrate horizon and below methane in gaseous form.

We see that in Erickson’s review the heat flux data were the subject of strong selection and corrections. Only deep boreholes can give reliable evidence of the Pleistocene-Holocene dynamic of ocean temperatures. All of them were unreasonably excluded from the analyses. On the other hand, clearly incorrect measurements, which didn’t show a decline in heat flux with depth, were given the highest reliability rate.

In the Erickson review only the very first boreholes were included. After that, hundreds of measurements were conducted, but the conclusions drawn in this review have never been questioned since.

Now let’s analyze the final review based upon all of the boreholes of the DSDP, up to leg 96 [Hyndman et al., 1987]. Based on these numerous data, its authors have drawn the following general conclusion: “there is no conclusive evidence of bottom water temperature changes of a few degrees or more over time scales of tens of thousands of years … this supports the initial hypothesis of Erickson et al. [1975]” (p.1573). In their review, the authors present a big table containing 86 sites with measurements of heat flow, which (in their opinion) support their conclusion. However, the authors note that “in this data summary we have included only sediment probe heat flow values having … at least three points with interpreted equilibrium temperatures that form a uniform gradient (or a constant heat flow with depth if the conductivity varies significantly). Some potentially good data have thus been excluded, but we feel that these are necessary criteria to establish the validity of a hole bottom temperature probe measurement” (p.1570). Thus, the authors at the first stage excluded a great number of good data only because they could be indicative of temperature changes of the ocean’s bottom. And after the selection they draw the conclusion that there is no proof of bottom water temperature changes. This is a very big methodological mistake. This analysis is incorrect.

If the authors have previously excluded from analyses all data that indicate temperature change on the ocean bottom then in their data summary among 86 sites should be many boreholes which approve the conclusion about stable temperature on the bottom. But firstly, the authors of the review stated that in this table only boreholes containing at least 3 measurement points were included. However, half of the sites which have got into the final table have only one measurement besides the bottom surface measurement. Based on these sites, the bottom water temperature dynamic can’t be judged. Two points can’t show profile bend. Secondly, 9 sites in the review table [Hyndman et al., 1987] are the boreholes from the isolated Mediterranean and Black Sea. Their temperature history is not connected with the open ocean. Thirdly, the majority of the remaining sites included in the table are boreholes shallower than 200 meters. They can’t be reliable evidence of the Pleistocene-Holocene ocean temperature dynamic.

We have found in the big sheet only 3 deep holes (>400 m) drilled in the open ocean in which three or more temperature measurements are present and there are thermal conductivity measurements (Figure 3). Of these, only one borehole is classified as having “good” quality (site 533), i.e. the site used by the authors to prove stability of heat flux with depth. However, as follows from the measurements data, and as was noted in the initial report , these data in fact show a strong decline in heat flux with depth (Figure 2B). And this isn’t the casual error. The authors of the review have seen that the heat flux is declining with depth and have tried to correct these data. They have reduced temperature at 156 m and 256 m to 0.6^{0}C. But this small correction does not change the situation notably. The thermal flow decreases with depth anyway. In addition, that correction is abusive as it was noted in the initial report that if there is a possibility of a correction, then it might only be correction toward an increase of temperature on the levels. Before the temperature probe was removed from the sediments, it hadn’t yet reached equilibrium and temperatures still continued to grow, so the researchers who conducted measurements accepted the maximum temperature reached.

Of the remaining 2 sites, graded as “fair”, site 582 also shows a very strong decline in heat flux with depth (Figure 2B). We should note that the maximum methane concentration there was discovered at a depth of ~300 meters. Likely, gas clathrates floor locates at this depth , and that means that the temperature there is +24^{o} C.

Site 397 was reported in the review as containing 3 temperature measurements. Even if only these three measurements were to be considered, heat flux would still decline with depth (figure 2B). But in fact, 5 measurements were originally made. The lower 2 measurements were excluded from the review analyses. But these are the most interesting measurements. Measurements at a depth of 1438 m gave a downhole temperature of 21-27 ^{0}C. The precision of these measurements is very low, but the fact that this measurement is close to being real value is supported by measurements at depths of 448 and 579 meters. They have high quality and show negative heat flow (Figure 2B). In Figure 1,B we see that this is possible at sharp temperature changes on the ocean’s floor at low geothermal flow. As it noted in the initial report, these sediments are rich with methane. The theoretical floor of gas clathrates is situated exactly at the bend in the temperature profile (Figure 2B). It is therefore likely that this represents a situation as presented in Figure 1B.

Finally, all three deep sites that were included in the review analyses to prove stable heat flux with depth in truth do not show this; they show a sharp decline of heat flux. We have restricted our analyses and have shown on the Fig. 2B,C only boreholes with depths >400 m, but a similar picture can be seen in data from shallower boreholes. For example, in the final table only 1 borehole (335) exists in the range 300-400 meters [Hyndman et al., 1987]. It has “excellent” quality on the sheet, i.e. it states that it reliably proves the cold ocean hypothesis [Hyndman et al., 1987]. But, as seen in the initial report , it also clearly shows a decline in conductive heat flux with depth. In this table all shallow boreholes from the Erickson report are included as proof of the stable heat flow, but as we showed earlier, all of them show heat flux decline with depth.

Now let's investigate the data of deep (>400 m) boreholes which were not included in the final table in the review, but were noted in the Appendix [Hyndman et al., 1987] (see Figure 2B). Besides the already mentioned site 185, there is also site 406 (Figure 3). It is presented in the table but only as a shallow borehole with 4 measurements. In truth, it was a deep borehole with 6 temperature measurements showing a decline of heat flux with depth (Figure 2b). At site 549 there was not only 2 measurements, as noted in the table, but 3, and there were heat flux declines there with depth as well. Site 568 was excluded from the table, but noted in the Appendix. There the heat fluxes declined also. The temperature profile of this borehole reached the bottom of gas clathrates (Figure 2b). This is validated by the drilling results. Gas clathrates were detected from this borehole at 190-315 m and 391-410 m depths (these gas clathrates were taken up on board). Below 410 meters, as drilling showed, gas is in a nonhydrated state. If 1 more temperature measurement below this depth had been conducted in this borehole then a sharp shift in the heat flux would likely have been obtained.

We checked all of the deep boreholes from the open ocean DSDP drilling program and found total discrepancy between the conclusion of the final review and the facts. Even careful data selection could not hide this. All the boreholes show that strong heat flow decreases with depth. There are no data proving the cold ocean hypothesis.

It should be noted that the authors of measurement have always seen heat flux decline with depth. The fact that the actual data contradict the accepted hypothesis has always troubled them. In most cases it is briefly noted in initial reports. But this topic is never deeply discussed.

Now let's investigate boreholes drilled in isolated basins – the Mediterranean and Black seas. There are 3 such deep boreholes contained in the table of the review: site 374 – 304 m, site 379 – 425 m, and site 380 – 370 m [Hyndman et al., 1987]. All of them have "excellent" quality. Their high quality is especially noted in the review [Hyndman et al., 1987]. They differ from boreholes in the open ocean by displaying stable heat flux. These measurements indicate that water temperatures in the bottom of these seas during the late Pleistocene differ not much from modern day (12.5°C and 9°C).

In reviews based on DSDP results [Erickson et al., 1975; Hyndman et al., 1987], deep temperature measurements were excluded from analysis. In following drilling projects such deep measurements became very rare. Among initial ODP and IODP reports we found only 4 deep boreholes in open ocean that have temperature measurements in undisturbed sediments deeper than 400 m. Let’s investigate these boreholes. What is interesting is that the temperatures in these boreholes were measured not only by temperature probe but with other methods too.

In a united 671-948 borehole, measurements were conducted during legs 110 and 156. The combined results are presented in Figure 2c. (We have corrected one obvious mistake made at temperature extrapolation at a depth of 247m.) The temperature gradient below 100 m stays almost stable and a decline in heat flux with depth occurs due to the reduction in heat conductivity. On the same site in parallel borehole 948D temperature loggers were installed. In 18 months, when borehole temperatures stabilized, they indicated approximately the same result – a stable temperature gradient (72°C/km) below 100 meters. This proves the reliability of the old technique.

At site 704 a decline in heat flux with depth was recorded (Figure 2c). Additionally, the temperature profile was calculated based on the porosity and resistivity of sediments at this site. It also showed a strong decrease in thermal flux with depth [Nobes et al., 1991].

On site 801 during leg 144, 2.5 years after drilling, the temperature profile was measured in detail. Data on the heat flux from this borehole are likely to be the most reliable of all. The decline in heat flux with depth found (Figure 2C) strongly correlates with model results (Figure 1A).

In borehole 1093 the heat flux also declined. Numerous measurements in the upper interval were conducted with limited distance between them, increasing errors in heat flux. If you average these data, then the decline of heat flux with depth becomes obvious. Importantly, although the measurement quality at the site was low, the most important measurement in this borehole (depth of 482 m) was made with “excellent” quality.

Borehole 1226 has crossed the line of stable gas clathrates at 305 m depth, and exactly at this point there is a sharp bend in temperature profile (Figure 2c). This may indicate the borehole piercing a low-power gas clathrates horizon. This is evidenced by high methane concentrations and a clear peak in the velocity of p-waves at this depth.

Now we investigate boreholes of isolated seas and shallow water. The Pleistocene-Holocene dynamic on the bottom should differ there from the open ocean bottom water. Borehole 1352 drilled on the continental border 60 km east of New Zealand to a depth of 344 meters is the only deep borehole which indicated heat flux increase with depth (Figure 2C). This means that in this region during the peak of glaciation, when New Zealand was covered with glaciers, water surface temperatures were close to zero (today ~9°C).

The Japan Sea freezes in the winter; it is connected to the ocean by only shallow straits. During legs 127 and 128, 5 boreholes were drilled to depths of 130 to 300 meters. All of them showed precisely uniform profiles of temperatures and stable heat flow. This means that bottom temperatures were stable (~0°C) both in glacial periods and today.

Borehole 1324 was drilled to the north of the Mexican Gulf at a depth of 1057 m. This site displays complicated bottom water temperature dynamics. In the upper part of this borehole a temperature gradient of ~100°C/km was measured, which then declined to 18.6-21.3°C/km. Deeper than 300 m it declined further to 16.2-16.7°C/km, and below 530 m it will likely rise again. Today at a depth of 1 km in the North Atlantic, water temperatures are 6-10°C, largely from Mediterranean Sea waters [Hay, 1983]. But the temperature at the ocean’s bottom of site 1324 is +2°C, meaning that there is an influence of water from Arctic seas. Possibly, initial Holocene bottom waters at this site were cold, then they were penetrated by Mediterranean waters before Arctic waters penetrated once again.

We have analysed all of the temperature measurements from sediments of deep oceanic boreholes. However, boreholes drilled in young ocean floor basalts also exist. Unfortunately, the top layer of young basalts is highly permeable and when they are pierced with a borehole, the flow of cold heavy water inside the borehole takes place [Hyndman et al., 1987]. Even a weak flow significantly changes the temperature profile: it becomes concave and smooth; it does not have bends which have to exist in places of sharp thermal conductivity changes (for example, site 1309D).

The most intensively studied basalt borehole is 504. It is the deepest borehole (2111 m) and contains detailed heat-conductivity measurements as well as permeability measurements. In the upper part of basalts, permeability is 10^{-13} m^2, and water flows into the borehole there. By 536 m depth, permeability decreases to 10^{-17} m^2 and water filtration is impossible below this point. These basalts are 5.9 million years old. On top they are covered with a 275-metre-thick layer of sediments. The heat flux measured in these sediments is equal to 196 mW/m^2. This corresponds with measurements made on bottom surfaces conducted in the area around this borehole. High-precision temperature measurements were conducted in this borehole during legs 92, 111, 137, 140 and 148. All measurements indicated that the conductive heat flux in basalts calculated through a temperature curve declined from 180-200 mW/m^2 at 550 m to 120-125 mW/m^2 at 1000-metres depth, and remained the same down to the bottom of the borehole. This is a very strong and reliable measurement of heat flux decline with depth, therefore several duplicating measurements were conducted, but all of them have indicated the same thing. All possible reasons for such a strong decline in heat flux (except for changes of bottom temperatures) were considered in detail (Initial Report, Leg 111). But a convincing reason was not provided. One potential explanation suggested that heat conductivity in basalts can increase with increasing pressure and temperature, and for values measured on board the ship should be made a correction for *in situ* conditions (leg 92). However, special investigations showed that this explanation doesn’t help – pressure influence is minor, and with temperature increase heat conductivity within the temperature range 28-170°C does not increase but declines by 0.0054-0.01 W/m °C for each degree of temperature rise [Schloessin and Drovak, 1997]. This phenomenon is more likely a general rule for crystal geological material; the only exclusion from this rule is glassy (amorphic) basalts [Petrunin et al., 1971; Schloessin and Drovak, 1997]. Experiments on sedimentary samples from hole 549 (calcareous silty mudstone, calcareous sandy mudstone, sandy limestone and sandy siltstone) indicate a significant decrease in conductivity also by between 0.007 and 0.011 W/m°C per degree over the temperature range zero to 80°C [Foucher et al., 1984].

For deep and hot boreholes it is a very strong correction; because of it, the heat-conductivity coefficient and calculated heat flux can be reduced twice. For deep boreholes in unconsolidated sediments this correction is also valuable, since sediments compress with depth. Analysing data from boreholes, this important correction was never applied, since it just strengthens the heat flux decline with depth, which, it is broadly considered, cannot happen. In contrast to mineral carcass, thermal conductivity of interstitial water increases with temperature rise [Erickson et al., 1975]. The correction mitigates thermal flow decrease by depth. Therefore it has always been done in full strength [Erickson et al., 1975; Hyndman et al., 1987], even when sediments' porosity and temperature gradient strongly decrease with depth.

3. Results

In the final reports it was announced that heat flow on the bottom of all oceans does not change with depth and therefore temperature on the ocean floor didn't differ much in the Pleistocene [Erickson et al., 1975; Hyndman et al., 1987]. But we have checked all deep open ocean boreholes and did not find evidence of that. In contrast, a sharp decline was observed in all holes (Figure 2). We can see the decrease of heat flow in the shallow boreholes and in the upper parts of the deep boreholes too (Figure 2). In boreholes 504, 801 and 948 temperatures were measured precisely (in detail) inside the borehole after they had reached heat equilibrium, and in all of these boreholes a strong decline of heat flux with depth was observed. All deep boreholes of the open ocean have indicated similar declines in heat flux with depth. We see that heat fluxes on the ocean floor are at least 40mW/m^2 higher than geothermal flux. If there is a bend in a temperature profile, then it corresponds with the boundary of stable gas clathrates (184, 397 and 1226) (Figure 2). Such results could not be just chance. In isolated seas, in which the Pleistocene-Holocene temperature dynamic was minor, heat flux stays constant with depth, and in shallow waters, where in the LGM water temperature was lower than today, heat flux grows with depth. This means that most temperature measurements were in truth done reliably and reflect the actual situation. These data give much interesting information.

We reached our conclusion of a heat flux decline with depth not by using statistical methods or averaging data sets, but from analysing each borehole. The conclusion of a 25°C temperature change in Greenland was based upon evidence from one core, yet we base our conclusion on numerous cores and all of them without a single exclusion indicated that the ocean floor in the LGM was warm. Subjective factors had a minor effect on our conclusions. We didn't conduct any measurements ourselves, taking all data from initial reports. The authors of these measurements were sure that heat flux is stable with depth. The subjective factor favored the cold ocean hypothesis.

The heat flux profile in bottom sediments is the most simple and reliable paleothermometer. It is based on a simple physical law – Fourier's law: heat flux is proportional to temperature gradient. We have tested the reliability of this paleothermometer on the ocean floor, in the Mediterranean, Black and Japan seas, and on the New Zealand shelf. It works everywhere reliably. We have confidence in it. The temperature profile shows the temperature itself – heat that is preserved in the sediments from the previous warming. All other paleothermometers are not connected with temperature directly. They are analysing some parameters which are connected with temperature.

These are empirical dependencies, and there is no confidence that these dependencies account for all factors.

At the moment we are not trying to reconstruct the last ocean cooling in great detail. Did cold water penetrate the Atlantic and Pacific interiors simultaneously? When did the strongest overturning occur, in Bølling-Allerød or after Younger Dryas? Yet, by comparing measured temperature profiles (and especially the profile of site 801) (Figure 2) with modelled ones (Figure 1) we see that ocean floor heating was strong, 25-30°C. Only such a heating would cause gas clathrates to melt, and at their second freezing sharp bends would occur in temperature profiles.

3.1. Geothermal flow

In theory, as the lithosphere cools (moving away from rift zones), heat flux must decline proportionally to the square root of crust age [Stein and Stein, 1992; Pollak et al, 1993]. As age increases by 4, heat fluxes decrease by 2, but in reality the decrease is very small. Today on average heat fluxes measured on the bottom surface through the ocean floor of the Miocene age (range 5.3-23.7 Myr) are equal to 81.9 mW/m^2, of the Oligocene age [23.7-36.6 Myr) to 62.3 mW/m^2 [Pollak et al., 1993]. On the oldest ocean floor (Late Jurassic, 144-163 Myr) the mean heat flux is relatively high, 51.3 mW/m^2 [Pollak et al., 1993].

In the first approximation, the heat flux through the old lithosphere is equal to the heat-conductivity coefficient of the lithosphere multiplied by the temperature gradient in the lithosphere, which in turn is equal to the change in temperature between the top and floor of the lithosphere, divided by its thickness. Consequently, to explain such high heat fluxes from the old ocean floor it is necessary to suppose very high solidus upper mantle temperatures of -1450°C (that is almost liquidus), and a very high coefficient of heat conductivity – 3.14 W/m$^{\circ}$C [Khromov and Petrosyantc, 2001]. Even at these high values, and with a lithosphere thickness of 95 km [Khromov and Petrosyantc, 2001], the heat flux will be only 48 mW/m^2 [Khromov and Petrosyantc, 2001].

However, using such parameters, heat fluxes calculated for the young oceanic lithosphere strongly exceed heat fluxes measured on the ocean bottom [Stein and Stein, 1992; Pollak et al., 1993] – for Miocene crust by 40 mW/m^2 and for Oligocene crust by 31 mW/m^2 [Pollak et al., 1993] – and parameters of the model are significantly overstated. But if we subtract 40mW/m^2 from these heat fluxes related to warmer oceans in the LGM, then deep heat flux declines with motion from the rift zone would be much higher: the Miocene crust would emit 41.9 mW/m^2, and the late Jurassic crust would emit 11.3 mW/m^2 – in four times. Same with the theory. If we take a more realistic value of solidus hydrous peridotites – 950°C [Nobes et al., 1991] – and a coefficient of heat conductivity of 1.5 W/m$^\circ$C (remembering that heat conductivity tends to strongly decline as temperatures increase [Schloessin and Drovak, 1977; Petrunin et al., 1971]), then for the same thickness of lithosphere we would obtain a heat flux of 15 mW/m^2, which is 3 times less. If we add 40 mW/m^2 to it then we would obtain a heat flux through the ocean floor of 55 mW/m^2. This is the most typical heat flux for ocean bottom [Pollak et al., 1993].

Estimating average heat flux through the ocean floor, a value of 101 mW/m^2 was obtained [Pollak et al., 1993]. At these heat flux calculations, fluxes through the young ocean crust (younger than 66 million years) were calculated by a model [Stein and Stein, 1992]. This model, as we showed, overestimated heat flux by ~ 3 times. On the rest of the territory, flux is overestimated by 40 mW/m^2 and possibly more, i.e by 3-4 times. As a result, average depth heat flux through ocean crust does not exceed 35 mW/m^2.

3.2. Features of thermohaline circulation in the ocean

Today, water in the ocean is stratified. The ocean interior is filled with cold and dense water, while the surface layer is filled with warm, and therefore lighter, water. The ocean is unstratified only in the high-latitude region, where both deep and surface waters are cold and dense. Such an ocean can be mixed to its entire depth solely with winds. Let us assume that wind is pushing the surface layer of water from the Northern Atlantic to the south, closer to the Antarctic. There, warm water is cooled and becomes almost as dense as on the ocean floor. Even with minor force this water can be submerged to the bottom. In the Northern Atlantic, water is transported south along the surface, and in exchange, deep water is emerging. This motion does not require a large effort—both on the surface and at depth, the water is similar in both temperature and density. As a result, under pressure from a “wind pump”, water in the conveyer belt will circulate from the surface of one cold sea to another.

Now let us investigate another hypothetical example: wind currents are absent, warm seas with high salinity are absent, geothermal heat sources are absent, and all water in the ocean is mixed and at a temperature of 20°C. In the mid- and high latitudes, water on the surface will cool and become heavier than the interior water. Strong thermal convection will take place: cold and dense water will submerge, and lighter, warmer water will emerge. The ocean will quickly cool and the territory where convection takes place will shrink. When the ocean temperature declines to +5°C, convection will take place only in the regions where surface water temperature is below +5°C. In time, the ocean will cool below 0°C. On the ocean floor, only water from the freezing seas will descend, and later only from the most cold and salty sea. As a result, the ocean will become filled with the coldest and most dense water. After that, convection will basically stop. Everywhere on the ocean, surface water will be lighter than water in the depth.

Now let us add to this scenario warm seas with high evaporation rates, with outflow of very salty and dense water. Until all the ocean is warm and strong thermal convection takes place; these small salty streams won’t be noticeable. But with time the ocean will cool, thermal convection will reduce, and these “haline” pumps will stop thermal convection. These seas absorb water with normal salinity from the ocean surface and return (by precipitation) fresh water, transporting salt down to the ocean interior. These seas desalinate the ocean surface. This reduces the density of the water on the surface and consequently stops high-latitude convection. Only haline pumps will continue working afterwards. They will strongly reduce salinity on the surface, and consequently water entering these seas will be fresher, and outflow will also become less salty and less dense. At some point, the haline pump will begin idling – outflow water will be lighter than interior ocean water, and outflow will be spread in the lower parts of the ocean surface layer. In the end, only the strongest haline pump will continue supplying water to the ocean bottom, but eventually it will stop. Thermohaline circulation will stop.

Thermohaline circulation appears only under buoyancy forces, i.e. upwelling water must be lighter than downwelling water. If their density becomes equal, motion forces will disappear and convection will stop. Convection takes place either when the ocean is cooling or when there are heat sources present at the bottom. Many researchers note that geothermal heat flow is important for oceanic circulation [see refs. in Adkins et al., 2005]. We especially note that with no heating from below, stable thermohaline circulation is impossible. The ocean water could be mixed by wind, but this is not thermohaline circulation. Rayleigh's formula describes conditions for thermal convection. In this formula, the thickness of the liquid layer is to a power of 3. For instance, in shallow lakes convection appears at several degrees of temperature difference between the surface and the bottom, whereas convection in the non-stratified ocean, which is 1000 times thicker, appears at a billion times less temperature difference. An energy value of 50 mW/m^2 is tremendous, enough even for convection to occur in the hard mantle. If all the energy converts into the kinetic energy of ocean currents then all ocean water could reach a speed of 0.7 m/sec in one year.

Now let us add to our hypothetical scenario geothermal heat flux (unrealistically strong to start with). First, an ocean at 20°C will cool, but then its temperature will stabilize: heat losses on the surface will be compensated for by heat flux at the bottom. Convection will be very strong and stable. If the geothermal heat flux is reduced, ocean temperature and thermal convection will also reduce correspondingly. Reducing the geothermal heat flux, we can reduce the convection to the point where the effect from haline pumps will become noticeable. At this point, they will quickly desalinate the ocean surface. Later, they will start idling and ocean ventilation will stop, but heating of the bottom waters will continue. Interior water temperatures will slowly increase. Water density will decrease until it becomes lower than the density of the water flowing out of the strongest haline pump. At this point, this pump starts working again. This pump will quickly reduce the surface salinity and will start idling again. Then the ocean will warm up again, and so on.

Glaciers have a large influence on haline pumps. Expansion of glaciers leads to increased salinity on the ocean surface and, consequently, increased salinity and density of water in haline pumps. Therefore as long as ice accumulates haline pumps will rarely idle. Conversely, if glaciers retreat, the surface salinity will decrease and haline pumps will idle until interior waters are heated.

Ocean heating cannot last forever. When at some point the density of water in the ocean depths will become less than in the cold sea, then thermo pumps will turn on, salinity on the ocean surface will immediately increase, and in cold seas very strong downwelling will occur (thus sharply warming Arctic climate). It will continue until ocean would release all accumulated heat. However, if there is a lot of ice on the surface of the cold sea and on surrounding land, then its quick melting will decrease salinity and water density in cold seas; the thermo pump will stop and the release of interior ocean heat will stop.

If, owing to tectonics or ocean regression, a strait which connects a haline sea with the ocean becomes narrower and/or shallower, then regardless of ocean surface salinity, the salinity of down flow from the haline pump will be high and decline in the salinity of the ocean surface will continue. In an extreme case, a strait can become so shallow that outflow from it will stop and an evaporating basin will occur. Salty water will still flow into the basin, and fresh water will return to the ocean surface with precipitation. Such a basin of the same size as the Red or Mediterranean Seas can pump out almost all the salt from the ocean surface, and the ocean will become ultra-stratified – water exchange between the surface and the interior ocean would be very low owing only to diffusion.

Imagine an idealised extreme case: the ocean surface is totally desalinated, the density of the surface water is close to 1000 kg/m^3, the interior ocean water salinity is $35^0/_{00}$ and diffusion exchange of salt and heat between the surface and the interior is absent. To take the ocean out of this stable state would require heating the interior water up to 77°C. Only at that temperature would the density of deep water drop to 1 g/cm^3 and convection would start.

In reality, as the ocean warms up, diffusion heat exchange with surface waters will increase and become equal to deep heat flux, and after that ocean temperature rise will attenuate. A stable state can then occur: fresh light water will lie above heavier but warmer water, i.e. because of diffusive heat losses, bottom waters will not be able to heat to the point of overturning. With diffusion, salt from the interior ocean will slowly penetrate to the surface. A salty sea will continuously pump out this salt. This condition will be stable until the haline pump significantly desalinates the entire ocean. Such a scenario likely did not occur in the Pleistocene, but it could have occurred in other geological epochs when big evaporating basins were formed.

4. Methods.

4.1. Mathematical model of thermohaline circulation

To illustrate these features of thermohaline circulation and to reconstruct its dynamic in the Pleistocene we present a simple 4-box model of the ocean. In this model we combine all cold seas into one box with an area of 10.8*10^6 km^2, which is eight times bigger than the area of the Norwegian Sea. The thickness of the top, well-ventilated layer in this sea (in this box) and ocean (in surface ocean box) is accepted to be 200 meters. The salty sea area is accepted to be 0.45*10^6 km^2, with an average depth of 550 meters. These values are the same as for the Red Sea. Thus we consider a situation where there is only one haline pump against all thermo pumps.

The energy supplied from the sun and atmosphere is accepted to be 65*10^8 J/m^2/yr to the high-latitude sea, 220*10^8 J/m^2/yr to the salty sea, and 130*10^8 J/m^2/yr to the box of the ocean surface [Khromov and Petrosyantc, 2001]. The surface of each of these boxes emits energy into the space, depending on the temperature of the water in this box, according to the Stefan-Boltzmann law [Khromov and Petrosyantc, 2001]. Part of the energy in the salty sea is spent on the evaporation of 3.3 meters of water annually. This water and the energy of phase change are transported to the surface of the open ocean box. The deep ocean box is receiving additional energy of 75 mW/m^2 from geothermal and biological heat flux.

We have accepted that the exchange of water between boxes is proportional to the difference in water density in these boxes, i.e. in this scenario we are considering only thermohaline circulation. The density of water in each box is calculated using a known formula based on salinity and temperature. The coefficient of water exchange between the salty sea and the ocean surface is accepted to be $0.003*10^{15}$ $m^6/kg*yr$. At this coefficient, salinity, temperature and water exchange in this sea under the present conditions will be the same as in the Red Sea. As long as the water density in these boxes is low, they exchange water only from the ocean surface box, but as soon as their density becomes more than in the bottom reservoir, all water flowing out from the salty sea drops down to the depths.

The coefficient of water exchange between the cold sea and the ocean surface box is accepted to be equal to $0.15*10^{15}$ $m^6/kg*yr$. As soon as the water density in the cold sea becomes higher than in the deep reservoir water starts to sink down. For this flow, we accept that the coefficient is equal to $5*10^{15}$ $m^6/kg*yr$. These coefficients are selected so that the temperature in the cold sea under the present conditions is close to 0°C and flux to the bottom is close to $0.6*10^6$ km^3/yr (~20 Sv). At the appearance of flow from the cold sea to the bottom, additional (compensative) flow appears from the ocean surface to the cold sea and from depth to the ocean surface. If the density of water in the surface ocean box becomes more than that at the ocean bottom then downwelling also occurs. For this water exchange we took the same coefficient as for the cold sea. As soon as the water density in one of the surface boxes becomes less than the water density in the bottom reservoir, downflows from this box stop.

All these currents redistribute salt and heat between boxes. As a result, for each box we have budget equations for water, salt and heat. Visual description of the model is presented in the figure 4.

Now let us discuss the wind's role in ocean mixing. The role of the wind in the water exchange in the Bab el Mandeb Strait can be neglected. Thermohaline circulation dominates there. But in the water exchange between ocean surface and cold sea this is a valuable parameter. In the first approximation, water exchange is proportional to wind strength, and wind strength is, in turn, proportional to the meridional temperature gradient of the atmosphere, which in its turn depends on the temperature difference between warm ocean and cold seas; i.e. we can accept this water exchange as being proportional to the temperature difference between ocean and cold sea. We accept that thermohaline circulation in these boxes is proportional to water density gradient. However, water density in these boxes is proportional to water temperature (salinity in these boxes is almost the same), so it appears that both wind and density mixing are proportional to temperature difference. Both these dependencies can be united using a united coefficient of proportionality. We can assume that the accepted above coefficient characterises both wind and density water exchange. The density of water in the warm ocean is substantially lower than in the interior ocean; therefore wind and diffusion exchange between these two boxes can be neglected. Wind exchange of water can occur between the cold and interior boxes. In modern models, such water exchange is obtained using accepted parameterization parameters. But in these seas heating from the bottom is not accounted for. This heating can sustain the observed water exchange, and without wind. We have a way to estimate the wind's role in the exchange with the bottom water. We know that the ocean bottom in the LGM was warm. All heat flow measurements in the sediments on the ocean floor show this. If wind exchange intensity were stronger than thermal convection, then haline pumps would not be able to freshen the ocean surface and turn thermal pumps off. If low power sources have managed to warm the ocean then wind couldn't stop it, and this means that wind exchange is weaker than thermal convection.

In our scenario, we have added the ice accumulation into the equations for cold sea. As soon as the average temperature of the cold sea drops below 0°C, fresh ice starts to accumulate both on the surrounding land and in the sea. Salinity in this box consequently increases, and an additional source of heat appears (80 calories per gram of frozen water). On the other hand, as soon as the temperature of the water in this sea increases above 0°C, ice which has accumulated on the coast or is floating in the sea melts, consuming energy and reducing salinity. We have accepted that ice build-up and melting is proportional to the change between 0°C and the temperature of water in the cold sea. Coefficients are such that a temperature change in the sea of -1°C in a thousand years accumulates $4.86*10^6$ km^3 of ice. We have also accepted that the ice accumulation rate decreases linearly as its volume increases (larger glaciers have greater deflation and ablation). We have accepted a coefficient of proportionality such that the ice volume could not exceed a value equivalent to a 150 m layer of ocean water. We have also accepted that area and depth of seas are not influenced by change in ice volume, only by deep reservoir volume change.

The depth of Bab el Mandeb strait was changing with changing ice sheet dynamics. However, in this scenario we have decided not to account for that. First of all, it was most important for us to see whether the modern Red Sea can turn all thermal pumps off. Secondly, at a four-fold decline in the cross section of this strait, because of increased salinity, water exchange there would decrease only two-fold. Salinity will increase and therefore the power of this freshening pump will not decline. The energy of this pump (the essence of its work) comes from the evaporation of 3.3 meters of water annually. This value is independent of the strait's cross section.

Saltier water is less compressible; thus it is harder to submerge it to the bottom. To describe this thermobaric effect and model its short-period cyclicity [Adkins et al., 2005] we use a restricted haline pump – water from this pump reaches a bottom reservoir only if its density is 0.25 kg/m^3 more than the current density in the bottom reservoir. Initial parameters are set at 25°C water temperature and $34.7^{0}/_{00}$ salinity in all boxes.

The essence of our model is very simple. We have set box sizes close to real basins, we have set external energy sources (in this model they are stable), and we have set coefficients of the water exchange intensity. In reality these are the sizes of the openings connecting boxes – cross sections of straights. Between salty sea and ocean, this strait (Bab el Mandeb)—this opening—is the smallest. The strait between cold sea and the ocean is 50 times bigger (this is the width of the Northern Atlantic), and the area of the opening between cold sea and underlying interior ocean is 33.3 times more than that (depth of the "straight' is equal to ocean depth there), i.e. we have only set the geometry of the system and external sources. The model calculates everything else: salinity, temperature, density, water exchange. Water exchange between boxes is proportional to the density difference between them.

4.2. Model results

The dynamic of modelled ocean parameters based on the solutions of the set of all budget equations described above appears in Figure 5. We see that our model ocean has lost all initial heat in 16,000 years. Temperature in the depth decreased from 25°C to 1.6°C. At that point a noticeable decline in the salinity of surface waters was observed, and the haline pump quickly turned off thermal circulation – a glacial period had started. During glaciation, the haline pump was often idling and worked permanently only at the end of the glaciation period. The duration of the glacial period was 112,000 years, the same as the Pleistocene glacial periods. For the duration of the glacial period, ice accumulation in seas and on land was equivalent to 149.5 meters of water in the world ocean. The salinity on the ocean surface declined to $31^{0}/_{00}$ and on the sea bottom increased to 36.5‰. These correspond to estimations made for the LGM. The average temperature of water in the deep reservoir increased to 23.4° C and water density dropped to 1,024.9 kg/m^3 and became the same as in the cold sea. After that, active downwelling appeared in the cold sea. Owing to the sea strong warming, all ice quickly melted. During several hundred years, there were no winters on the coastal areas of the seas. In 14,600 years, the ocean has again lost all accumulated heat and a new glaciation period has started. We found that the duration of the interglacial was the same as the age of the sharp warming in Greenland in Boling-Alerd 14,650 years ago. Our model has shown a temperature of 1.6°C on the ocean bottom at the end of the interglacial period. This is the present ocean temperature.

We see that the simple model of thermohaline circulation under stable external conditions shows asymmetric cycles which look like glacial cycles. We did not try to make a super-precise model; all initial parameters are rounded values. However the duration and amplitude of both glacial cycles correspond with observed values. The model even indicates a short cooling somewhat similar to the Younger Dryas and also short-term cycles similar to Dansgaard/Oeschger warm events.

To test the model, we have substantially changed initial parameters and increased the number of boxes in the model, but if the depth heat flux is small and if the "pressure" of haline pumps is stronger than that of cold pumps then in all cases we obtain ocean heating. Glacial cycles get shorter if bottom heating is increased. If the haline pump is further restricted, this causes occasional turning on of the cold pump and Dansgaard/Oeschger warm events appear [Adkins et al., 2005]. If we remove glaciers from the model, the haline pump is often idling and glacial cycles are longer. But if deep heat flux were removed, then thermal circulation in the ocean would stop, and soon haline pumps would idle and thermohaline circulation would stop forever.

For the "Northern Atlantic" we have obtained a very strong temperature dynamic, but for the rest of the ocean the temperature dynamic on the surface does not exceed 1°C (see figure 1). But if were to add the Earth's orbit dynamic, decline in greenhouse gas concentration in the atmosphere, albedo change related to forest area decrease, and increase of snow and ice covered territories, then ocean surface cooling in the glaciation period would be substantially stronger.

Here we want to note that if a more detailed model were to be developed, it should take into account that the main ocean heating from the bottom occurs in the rift zones, on depths of 2–3 km, and big masses of water penetrating from the Red Sea at the beginning of glaciation would be cooling, heating up the cold floor.

The main proof of ocean warming in the last glaciation period is the high remaining heat flux from the sediments of the ocean floor. Our model has shown the mechanisms and dynamics of this warming. Close correlation between observed values and modelling results is an additional proof for ocean warming.

5. Discussion.

5.1. Consequences of warm ocean hypothesis

Many of the views on natural processes in the Pleistocene are coordinated with the cold ocean hypothesis. Rejecting this hypothesis requires a revision of these views.

Today, air temperature at the North Pole in summer is close to 0°C. At that time, ice melts there. The Arctic receives lots of heat from the Atlantic. Therefore ice cover in the Arctic Ocean is thin and firm. Ocean currents and wind break this cover and transport ice south. If the temperature in Greenland in the LGM was 25°C colder than it is today [Dahl-Jensen, 1998], and if the heat conveyer did not work, then ocean ice cover would have been thicker and would never have melted [Bradley and England, 2008]. Sea water freezing becomes slower as ice thickness grows. But the surface of Arctic ice receives precipitation of, on average, ~10 cm of water in the form of snow. During the glaciations, when the Arctic was substantially colder, this snow never melted. The areas of the Antarctic and Arctic Oceans are comparable. The amounts of snow precipitating on these areas are also comparable. In the Antarctic, the thicker the ice sheet the faster is its motion. Antarctic ice flows to the ocean through a "strait" almost 20,000 kilometers wide. In the Arctic Ocean, the situation is opposite: ice flows south through narrow and shallow straits. The thicker and harder is the ice in the Arctic, the harder it is for winds and currents to press it through these straits. In the glaciation period, these straights were even shallower. The smaller the ice outflow, the more intensively it accumulates in the Arctic. At such dependency, spontaneous and irreversible glaciation of the entire Arctic Ocean is possible. There must have been floating sea glaciers with very low ice $\delta^{18}O$. These glaciers would have been mostly submerged, with only 9% of their volume above the water surface, i.e. only 90 m of a 1 km deep glacier would be above sea level. Such a glacier could not flow into land or penetrate to the Atlantic through shallow straits. A 1 km iceberg could not go through a 100 m depth strait. At an ice accumulation rate of 10 cm/yr, the Arctic Ocean could be filled three times over with free-saline ice during glacial periods. If this glacier hit the bottom of the ocean somewhere, then it should start growing rapidly above water level in this place. This would result in the ice flowing.

In the global budget of $\delta^{18}O$, made in the framework of the cold ocean theory, there is no place for such a big floating reservoir of isotopically light ice. However in the framework of the warm ocean, the appearance of such a reservoir is very likely.

During the last glaciation, $\delta^{13}C$ in the ocean water declined by 0.3–0.4‰. It is assumed that the reason for this was that the terrestrial reservoir of organic carbon decreased by 500 Pg C, because of forest areas shrinking, and this isotopically light carbon was absorbed by the ocean [Sigman and Boyle, 2000]. The methane content in depth sediments is 10000–20000 Pg [Kennet et al., 2000]. Today it is stable; layers of solid gas clathrates prevent underlying methane from escaping. But if water temperature on the ocean bottom has changed significantly, then most of the gas clathrates of the World Ocean melted during glacial periods. Water level and pressure were lower, and this accompanied the emission of hundreds of pentagrams of methane from bottom sediments, dissolution of methane in the water, and further oxidation to CO_2. $\delta^{13}C$ of gas clathrates is very low (-65‰) [Kennet et al., 2000], therefore $\delta^{13}C$ of ocean water was reduced during glaciations. An emission of 170 Pg C equivalent of this methane (this is only 1–2% of global storage of the methane on the ocean bottom) will have the same effect on $\delta^{13}C$ of oceanic water as an emission of 500 Pg C from terrestrial or oceanic organic reservoirs.

At low circulation in glacial periods, oxygen input to the bottom decreased, but at the same time nutrition supply to the surface from the depth also declined by the same level of magnitude and consequently the biological pump was reduced as well, i.e. oxygen input was lower but its consumption was low as well. Therefore there were few anoxic conditions on the ocean floor. Additionally, bottom water was heated by geothermal and biological heat, therefore despite low content of dissolved gases, their partial pressure was increasing. If such warm gas-saturated water eventually emerged on the ocean surface, it would first emit gases into the atmosphere, and only after cooling would it strongly absorb them.

During the glaciation period CO_2 concentration in the atmosphere declined; it lost ~200 Pg of carbon. Carbon content in forests was reduced by 500 Pg. It is thought that this carbon was absorbed by the ocean and then in deglacial periods it was returned to the atmosphere and to the land [Sigman and Boyle, 2000]. However, mechanisms for this are still unclear [Sigman and Boyle, 2000; Kohfeld et al., 2005]. On the contrary, there are several data sets which give evidence of decreased carbon storage in the glacial ocean: i) Storage of organic carbon in the ocean is comparable in size with the terrestrial organic carbon reservoir (their $\delta^{13}C$ signatures is the same) [Brovkin et al., 2002]. Ocean productivity in the glacial period decreased significantly [Kohfeld et al., 2005; Anderson et al., 2009; Zimov et al., 2009]. Only at the end of glaciation, when the ocean had substantially heated and circulation had slightly increased (see Fig. 1), did productivity increase in the Atlantic [Kohfeld et al., 2005; Zimov et al., 2009]. Therefore the dissolved organic carbon content of the ocean's water and carbon storage at the bottom should be correspondingly lower; ii) The salinity of the glacial ocean increased (because of glacier growth), and its CO_2 solubility capacity decreased correspondingly [Sigman and Boyle, 2000); iii) An attempt to find proof of 700 Pg deglaciation emission from the ocean to the atmosphere was made by Yu et al. [2010]. The data indicated that the carbonate ion concentration of the deep ocean during deglaciation increased only by 10 mmol/kg, and only in the deepest part of ocean [Yu et al., 2010]. It is equal to a 100 Pg C loss from the deep reservoir [Yu et al., 2010]. In the other 70% of the ocean volume the carbonate ion concentration strongly decreased during deglaciation [Yu et al., 2010; Zeebe and Marchitto, 2010] (by 60 mmol/kg at the ocean’s surface [Yu et al., 2010]). If all these data are interpolated, then the carbonate ion content of the entire ocean decreased by 20–25 mmol/kg and the ocean absorbed as little as 700 Pg inorganic carbon during deglaciation.

CO_2 dissolution in water is strongly dependent on water temperature. Therefore carbon storage in the ocean can be increased by cooling of the ocean [Sigman and Boyle, 2000]. However, the possibilities of this explanation are limited, as today the interior ocean temperature is close to the point of freezing. Therefore only one third of atmospheric CO_2 decline in glaciation can be explained by slight cooling (2–3°C) of the ocean [Sigman and Boyle, 2000]. But all deep boreholes show strong ocean warming. The ocean was half-filled with warm (much CO_2 depleted) water from the Red Sea, and the carbon content of ocean water during the glaciations must have been very strongly reduced because of that. Accounting for all of the above, carbon storage in the LGM ocean was reduced by approximately 1,500 Pg C. That means that there should be some other big carbon reservoir which greatly increased in size during the glaciation period and absorbed carbon from the atmosphere, forest ecosystems, gas clathrates, and ocean carbon reservoirs. Of all known carbon reservoirs, only permafrost and soils of the mammoth steppe biome could absorb such amounts of carbon. In the LGM, resting on the permafrost mammoth steppe was the earth’s biggest biome. It was a highly productive ecosystem, where hundreds of millions of large herbivores maintained their pastures [Zimov et al., 2012]. Only soils of the cold biome are capable of accumulating hundreds of kilogram’s of carbon per square meter [Zimov et al., 2006; Zimov et al., 2009]. Today, permafrost is the largest reservoir of organic carbon (1670 Pg C [Tarnocai et al., 2009]), and in the LGM it was at least twice as big. The mammoth steppe biome pumped light carbon from other reservoirs into its frozen soils.

During the Pleistocene-Holocene transition increased not only atmospheric CO2 concentration, but also concentration of the atmospheric methane. An analysis of global ^{14}C data for basal peat combined with modelling of wetland succession allowed us to reconstruct the dynamics of global wetland methane emission through time. These data show that the rise of atmospheric methane concentrations during the Pleistocene-Holocene transition was not connected with wetland expansion, but rather started substantially later, only 9 thousand years ago. [Zimov and Zimov 2014]. The isotopic composition of methane varies according to source. Owing to ice sheet drilling programs past dynamics of atmospheric methane isotopic composition is now known. Modelling of the budget of the atmospheric methane and its isotopic composition allowed us to reconstruct the dynamics of all main methane sources. For the late Pleistocene, the largest methane source was megaherbivores, whose total biomass is estimated to have exceeded that of present-day humans and domestic animals. During deglaciation, the largest methane emissions originated from degrading frozen soils of the mammoth steppe biome. Methane from this source is unique, as it is depleted of all isotopes. We estimated that over the entire course of deglaciation (15,000 to 6,000 year before present), soils of the mammoth steppe released 300–550 Pg (10^{15} g) of methane. This is approximately 300 Pg C [Zimov, Zimov 2014].

What was the total carbon loss from the northern soils over that time period? Even in fully anaerobic conditions only part of the stored organic (28±12%) can be transformed into methane [Walter Anthony et al. 2014]. When thick (tens of meters) frozen ice-rich soils (yedoma and its southern analogues) thaw they are initially anaerobic, however, as permafrost degradation continues, underlying gravel and sand thaws and water begins to drain from shallower soil layers. These soils become aerobic and carbon is transformed into the CO2. However on most territories occupied by the mammoth steppe, soils (both including active layer and soils incorporated into the permafrost) were shallow (less than 2–3 meters deep), and carbon stored in these soils was mostly decomposed under aerobic conditions when the climate warmed. Therefore, if we accept that 15% of permafrost carbon loss was transformed into methane, we can estimate that permafrost soils lost 2000 Pg C during deglaciation (15–6 ka BP) [Zimov and Zimov 2014]. This carbon is sufficient to supply Holocene forest expansion and fill the cooling ocean with CO2.

At present, the ocean has already released most of the accumulated heat from the last glaciation. Bottom sediments have also cooled, so thermo circulation has decreased. Soon it can be expected that salty seas will stop thermo circulation and a new glacial period will start (see Fig. 1).

By artificially regulating water exchange in narrow straits which connect salty seas with the ocean, we may have the capacity to change the salinity and density of outflowing water. By increasing water exchange we might stop the freshening of surface water and may prolong the life of cold pumps. In contrast, if we could reduce the water exchange, the thermo circulation might slow more quickly and hasten the beginning of the next glacial cycle. This, to some extent, can compensate for warming caused by increased greenhouse gas concentrations in the atmosphere.

However, the results of this experiment are doubtful. Today, the ocean absorbs half of all anthropogenic emissions of CO_2. But if thermo circulation is replaced by haline circulation, then the ocean interior will start warming and releasing CO_2 in addition to anthropogenic emissions. In the past, this emission was totally compensated for by permafrost and mammoth steppe expansion, but with the anthropogenic rise in atmospheric CO_2 concentrations, new permafrost will not be formed, and northern steppes will not expand in Europe. As a result, by changing thermo circulation to haline circulation we would double the rate of emissions of atmospheric CO_2 and see a strong decline in ocean productivity.

6. Conclusions

Forty years ago, cold and warm ocean hypotheses were equal in rights – both were discussed. Neither one of them had enough proof. During that period of time the geology revolution started – the theory of plate tectonics was evolving, and it was very important to know why bottom sediment emits energy: either it is a mantle heat flow or these are bottom sediments releasing the heat accumulated during the last ocean warming. Temperature profiles from deep boreholes answer this question simply and reliably – if heat flux is stable with depth, then the first hypothesis is correct; if it declines, then the second hypothesis is correct. Oceans are connected with each other, therefore hypotheses are alternative: heat fluxes in the ocean bottom sediments should be either stable everywhere or decline everywhere. Therefore it is enough to reliably measure temperature in a few boreholes to choose the correct hypothesis.

First three profiles of heat flux in ocean sediments were obtained in boreholes 184,185 and 206. These are deep boreholes, and they have indicated a strong decline in heat flux with depth. Solely because of that, these data were excluded from previous analyses. All consequent deep boreholes indicated the same results and they were either also excluded from analyses or, contrary to the facts, it was stated that heat fluxes in these boreholes are stable and they prove the cold ocean hypothesis.

One of the reasons why data from deep boreholes were not trusted was that they indicated extremely strong ocean warming in the LGM. It wasn’t clear what the source for that ocean heating was. However, such sources exist, and they are sufficient for ocean warming. The heavy waters from the Red Sea, geothermal heat flux and decomposition of organic matter on the ocean floor. Flux of salty water to the ocean bottom from the Red Sea maintained freshening of the ocean surface and strong ocean stratification. In the absence of ocean mixing, even weak energy sources can warm the ocean interior. Strong stratification of water in the LGM is supported by data on salinity of the ocean on the surface and at the bottom, and data on low bioproductivity. The idea that the ocean in the glaciation was filled with Red Sea waters is supported by data on similar isotopic content of water in this sea and ocean bottom water in the LGM.

We have made a mathematical model of thermo-haline circulation of the world ocean. We have taken sun energy income to the ocean surface, high-latitude seas and the Red Sea as constant and the same as today. Geothermal and biogenic bottom water heating we also took as constant. But as a result we obtained strongly asymmetrical glacial-interglacial cycles of ocean dynamic, climate and glaciers. The model showed that during the glacial period ventilation of the ocean sharply diminished. Thermal pumps didn't work, high-latitude seas froze. Warm and heavy water from the Red Sea filled all interior oceans. In 112,000 years, the salinity of surface waters has declined to 31‰ (on the bottom it reached 36.5‰), and the average temperature of water in the interior ocean box reached 23.4^{o} C. At that, its density has become lower than the density of water in high-latitude seas, which caused the start of very strong convection. As a result, in 14,600 years, the ocean has fully released all heat accumulated in the glacial period, and cooled to 1.6^{o} C. Thermal pumps have, at that, slowed down, and haline pumps, at that, quickly decreased the salinity and fully stopped the thermal circulation. Ocean interior warming has started anew – a new glacial cycle has started anew. The cyclicity obtained in the model is very close to observed glacial-interglacial cycles, both by the amplitude and by the duration.

The uniqueness of the variable Pleistocene climate is likely connected with the fact that very cold and very salty seas and frozen soils, which could accumulate large amounts of carbon, appeared on our planet at the same time. From the analyses presented it follows that during glaciation epochs, haline circulation dominated – the ocean was taking up warm water from the surface and accumulating heat. On land, polar oceans and permafrost were accumulating ice and ocean bottom "ice" (gas clathrates) was melting. Interglacials are epochs when the ocean is dominated by thermo circulation – the ocean absorbs the coldest water and releases its heat. In the ocean bottom, water and methane crystallise while on the land, glaciers and permafrost thaw. Microbes turn into CO_2 and CH_4, the organic which is accumulated in the permafrost and cold soils.

Glaciations are periods when the ocean accumulated energy, and interglacials are periods when the ocean quickly released this energy.

References

1. Adkins, J.F., Mclntyre, K., and D.P. Schrag (2002), The salinity, temperature and $\delta^{18}O$ of the glacial deep ocean. Science, 298: 1769.
2. Adkins, J.F., Ingersoll, A., and C. Pasquero (2005), Rapid climate change and conditional instability of the glacial deep ocean from the thermobaric effect and geothermal heating. Quat Sci Rev, 24: 581.
3. Anderson, R.F., Ali, S., Bradtmiller, L.I., Nielsen, S.H.H., Fleisher, M.Q., et al. (2009), Wind-Driven Upwelling in the Southern Ocean and the Deglacial Rise in Atmospheric CO_2. Science 323: 1443.
4. Badawi, A., Schmiedl, G., and C. Hemleben (2005), Impact of late Quaternary environmental changes on deep-sea benthic foraminiferal faunas of the Red Sea. Marine Micropaleontology 58: 13-30.
5. Bradley, R.S., and J.H England (2008), The Younger Dryas and the sea of ancient ice. Quat Res 70, 1.
6. Brovkin, V., Hofmann, M., Bendtsen, J., and A. Ganopolski (2002), Ocean biology could control atmospheric d13C during glacial-interglacial cycle. GGG 3, 10.1029/2001GC000270.
7. Bryan, S.P., and T.M. Marchitto (2008), The Mg/Ca - temperature proxy in benthic foraminifera: New calibrations from the Florida Straits and a hypothesis regarding Mg/Li, Paleoceanography, 23: 1553.

8. Chapin, III F.S., Matson, P.A., and H.A. Mooney (2002), Principles of terrestrial ecosystem ecology. Springer-Verlag, New York.

9. Chuprynin, V.I. (1985), Interrupted autooscillation of geophysical systems. Nauka, Moscow,. (in Russian)

10. Cuffey, K.M., Clow, G.D., Alley, R.B., Stuiver, M., Waddington, E.D., and R.W. Saltus (1995), Large Arctic Temperature Change at the Wisconsin-Holocene Glacial Transition. Science 270: 455.

11. Dahl-Jensen, D., Mosegaard, K., Gundestrup, N., Clow, G.D., Johnsen, S.J., et al. (1998), Past Temperatures Directly from the Greenland Ice Sheet. Science 282: 268.

12. de Vernal, A., and C. Hillaire-Marcel (2000), Sea-ice cover, sea-surface salinity and halo-/thermocline structure of the northwest North Atlantic: modern versus full glacial conditions. Quat Sci Rev 19: 65.

13. Erickson, A.J., Von Herzen, R.P., Sclater, J.G., Girdler, R.W., Marshall, B.V., et al. (1975), Geothermal measurements in deep-sea drill holes. J Geophys Res 80*:* 2515–2528.

14. Foucher, J.P., Chenet, P.Y., Montadert, L., and J.M. Roux (1984), Geothermal measurements during Deep Sea Drilling Project leg 80. Initial Rep. Deep Sea Drill. Proj. 80: 423.

15. Geiselhart, S. (1998), Late Quaternary Paleoceanographic and paleoclimatic history of the Red Sea during the last 380,000 years: evidence from stable isotopes and faunal assemblages. Tüb Micropälaontol Mitt 17, 1.

16. Hay, W. (1983), The global significance of regional Mediterranean Neogenepaleoenvironmental studies. In: Meulenkamp JE editor. Bull. 30 Reconstruction of marine paleoenvironments. Loonzetterij Abe, Hoogeveen Netherlands.

17. Hemleben, C., Meischner, D., Zahn, R., Almogi-Labin, A., Erlenkeuser, H., et al. (1996), Three hundred eighty thousand year long stable isotope and faunal records from the Red Sea: Influence of global sea level change on hydrography. Paleoceanography 11: 147-156.

18. Hyndman, R.D., Langseth, M.G., and R.P. Von Herzen (1987), Deep Sea Drilling Project geothermal measurements: A review. Rev of Geophys 25, 8: 1563.

19. Kennett, J.P., Cannariato, K.G., Hendy, I.L., and R.J. Behl (2000), Carbon isotopic evidence for methane hydrate instability during quaternary interstadials. Science 288: 128.

20. Khromov, S.P., and M.A. Petrosyantc (2001), Meteorology and Climatology. Moscow University Press, Moscow.

21. Kohfeld, K.E., Le Quéré, C., Harrison, S.P., and R.F. Anderson (2005), Role of marine biology in glacial-interglacial CO2 cycles. Science 308: 74.

22. Luz, B., and Z. Reiss (1983), Stable carbon isotopes in Quaternary foraminifera from the Gulf of Aqaba (Elat), Red Sea. In: Meulenkamp JE editor. Bull. 30 Reconstruction of marine paleoenvironments. Loonzetterij Abe, Hoogeveen, Netherlands.

23. Miles, P.R. (1995), Potential distribution of methane hydrate beneath the European continental margins. Geophys Res Lett 22: 3179.

24. Nobes, D.C., Mienert, J., Mwenifumbo, C.J., and J.P. Blangy (1991), An estimate of the heat flow on the meteor rise, site 704. Proceedings of the Ocean Drilling Program, Scientific Results, 114: 39.

25. Petrunin, G.I., Yurchak, R.I., and G.F. Tkach (1971), Temperature conductivity of Basalts at temperatures from 300 to 1200°K. Izv: Earth Phys, 2: 65. (In Russian)

26. Pollak, H.N., Hurter, S.J., and J.R. Johnson (1993), Heat flow from the Earth's interior: Analysis of the global data set. Rev of Geophys 31: 267.

27. Sarnthein, M. (2011), Paleoclimate. Northern meltwater pulses, CO2, and changes in Atlantic convection. Science 331: 156.

28. Schloessin, H.H., and Z.D. Drovak (1977), Physical properties of samples from the joides, leg 36, Deep Sea Drilling Project. Initial Rep Deep Sea Drill Proj 37: 403.

29. Severinghaus, J.P., and E.J.Brook (1999), Abrupt climate change at the end of the last glacial period inferred from trapped air in polar ice. Science 286: 930, 1999.

30. Sigman, D.M., and E.A. Boyle (2000), Glacial/interglacial variations in atmospheric carbon dioxide. Nature 407: 859.

31. Stein, C.A., and S. Stein (1992), A model for the global variation in oceanic depth and heat flow with lithospheric age. Nature 359: 123.

32. Tarnocai, C., Canadell, J.G., Schuur, E.A.G., Kuhry, P., Mazhitova, G., and S. Zimov (2009), Soil organic carbon pools in the northern circumpolar permafrost region. Global Biogeochem Cycles 23: 2023.

33. Van der Zwaan, G.J., Duijnstee, I.A.P., Den Dulk, M., Ernst, S.R., Jannink, N.T., et al. (1999), Benthic foraminifers: proxies or problems? A review of paleocological concepts. Earth Sci Rev 46: 213–236.

34. Walter Anthony K.M., Zimov S.A., Grosse G., Jones M.C., Anthony P, et al. (2014), Permafrost thaw by deep lakes: from a methane source to a Holocene carbon sink. Nature, Submitted.

35. Washburn, A.L. (1979), Geocryology, A survey of periglacial processes and environments. Edward Arnold, London.

36. Yershov, E.D. (2004), General Geocryology. Cambridge University Press.

37. Yu, J., and H. Elderfield (2008), Mg/Ca in the benthic foraminifera *Cibicidoides wuellerstorfi* and *Cibicidoides mundulus*: Temperature versus carbonate ion saturation. Earth and Planet Sci Lett 276: 129.

38. Yu, J., Broecker, W.S., Elderfield, H., Jin, Z., McManus, J., and F. Zhang (2010), Loss of carbon from the deep sea since the Last Glacial Maximum. Science 330: 1084.

39. Zeebe, R.E., and T.M.Marchitto, Jr. (2010), Atmosphere and ocean chemistry. Nature Geoscience 3: 386.

40. Zimov, N.S., Zimov, S.A., Zimova, A.E., Zimova, G.M., Chuprynin, V.I., and F.S. III Chapin (2009), Carbon storage in permafrost and soils of the mammoth tundra-steppe biome:

role in the global carbon budget. Geophys Res Lett 36, L02502, doi:10.1029/2008GL036332.

41. Zimov S.A., and N.S. Zimov (2014), Role of Megafauna and Frozen Soil in the Atmospheric CH_4 Dynamics. PLoS ONE 9(4): e93331. doi:10.1371/journal.pone.0093331.
42. Zimov, S.A., Zimov, N.S., Tikhonov, A.N., and F.S. III Chapin (2012), Mammoth steppe: a high-productivity phenomenon. Quat Sci Rev 57: 26-45.
43. Zimov, S.A., Schuur, E.A.G., and F.S. III Chapin (2006), Permafrost and the global carbon budget. Science 312: 1612.

Figure captions

Figure 1. Profile of bottom sediment temperatures and profile of conductive heat flows for scenarios of cycling changes of ocean bottom temperatures from $+2^0$C to $+32^0$C during glacial-interglacial dynamic. The red lines are temperature profiles for various deep heat flows at the time of maximum heating of the ocean floor. The green and blue lines are temperature and heat flow profiles for the time periods of 11470 years and 14450 years after sharp ocean cooling. A –scenario for sediments not saturated with methane. B –scenario showing gas hydrate formation. Knees on the temperature profiles and leaps on heat flow profiles show gas hydrate bottom. C – temperatures and heat flows at stable temperature on the bottom ($+2^0$C) and at the same deep heat flows.

Figure 2. Profiles of temperatures, thermal conductivity, temperature gradients and heat flow in the bottom sediments in the open ocean.

A – boreholes measured during legs 1-26 and available in the review [Erickson et al., 1975]. Data marked with orange columns were taken from the resultant graph [Erickson et al., 1975]. The blue arrows are our data corrections. B – all deep boreholes measured during DSDP and noted in the review (Hyndman et al., 1987). C – deep boreholes measured during ODP and IODP. Thermal conductivity data: large dots – are average values for depth intervals as shown in the initial reports; small dots are initial values; circles are average values of thermal conductivity per depth intervals that we calculated from initial data. The numerals on the figures are number of leg, number of borehole and ocean depth. The blue dotted lines are the depth where gas hydrates freeze or thaw [Miles, 1995]. If sediment temperature lies to the left of this line, then at high methane concentration it will be in the form of gas clathrates. To the right of this line gas clathrates are absent. All boreholes in the figure show a strong decrease of heat flux with depth. This means strong heating of the ocean in the LGM. Only borehole 1352 (depth 344 m) shows an increase in heat flow with depth. This states that on the New Zealand shelf water in the LGM was colder than today.

Figure 3. Map of boreholes noted in the Figure 2.

Figure 4. Ocean circulation scheme for glacial and interglacial used in presented model.

Figure 5. Results of model of thermo-haline circulation in the World Ocean. Red line – Red sea; Blue line – high latitude seas; Green line – Ocean surface; Black line – Ocean interior.

Fig. 1

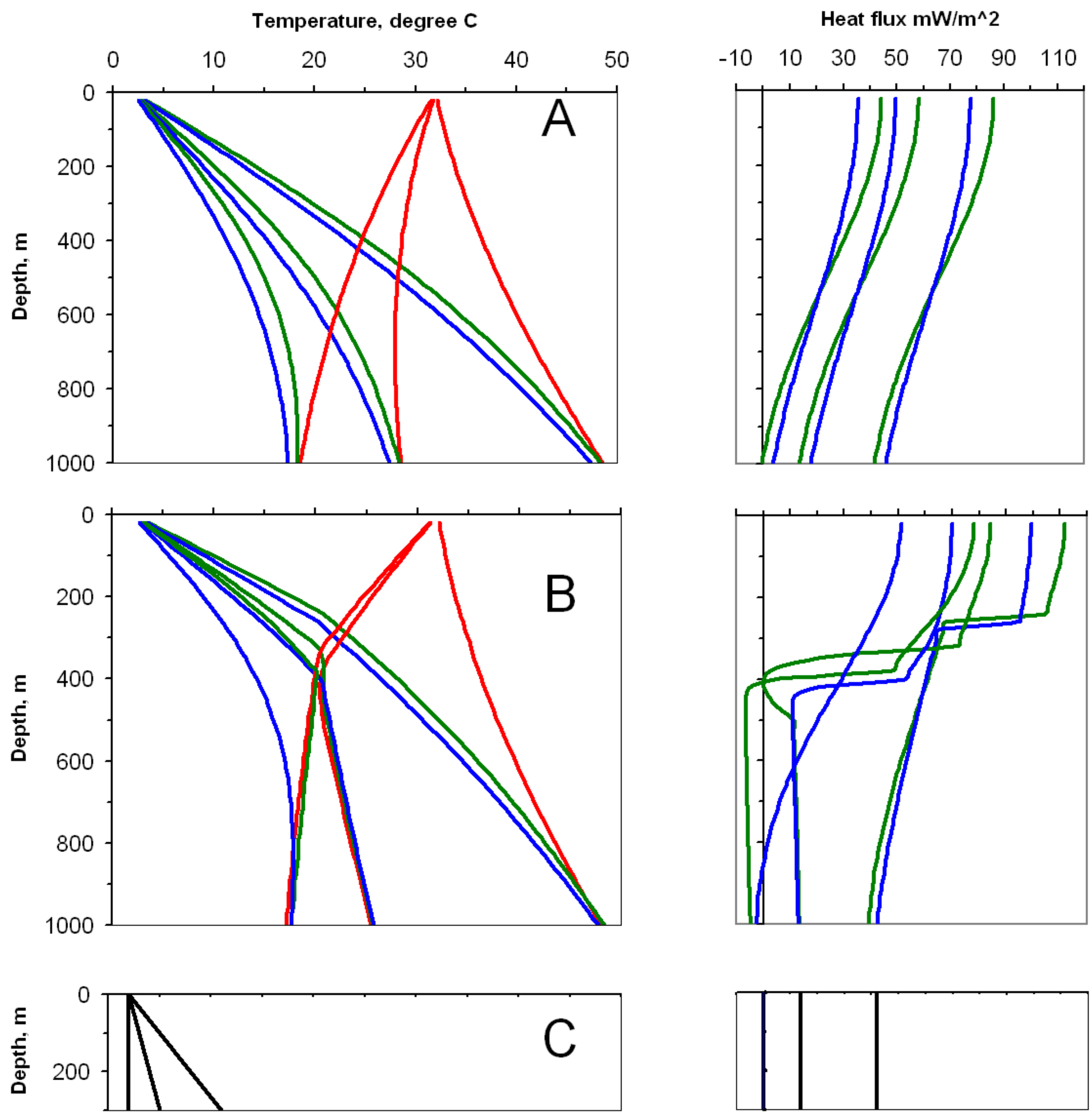
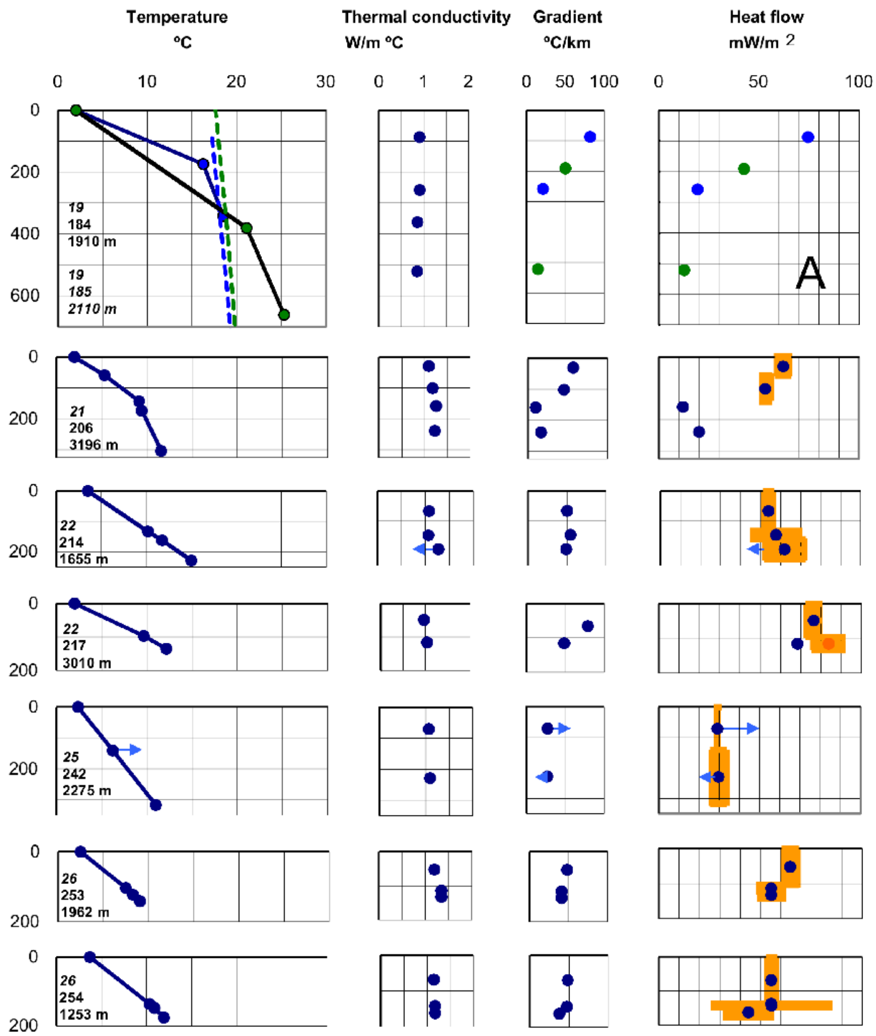
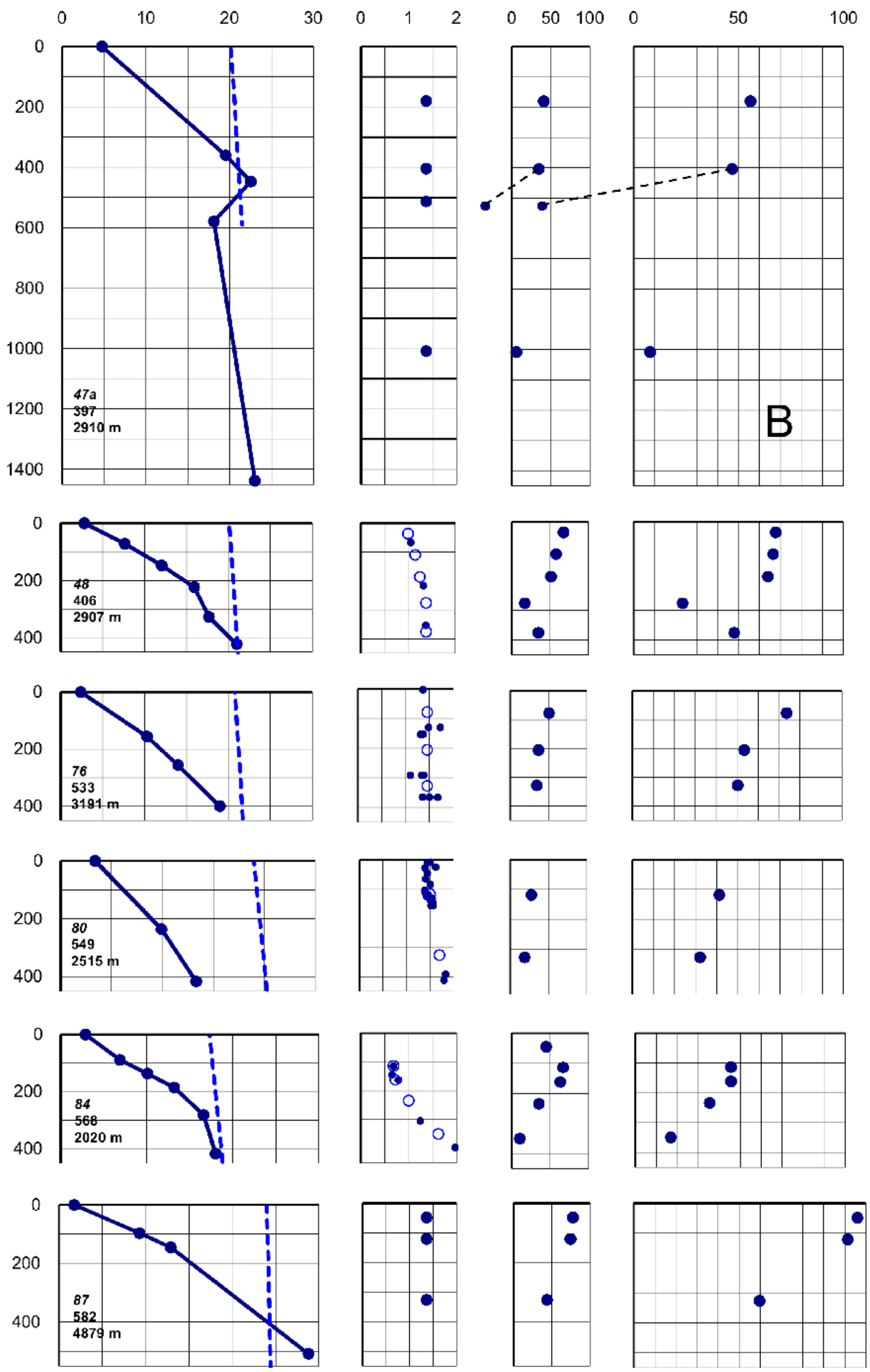

Temperature, degree C
0
10
20
30
40
50
Heat flux mW/m^2
-10
10
30
50
70
90
110
Depth, m
0
200
400
600
800
1000
A
B
C

Fig. 2a

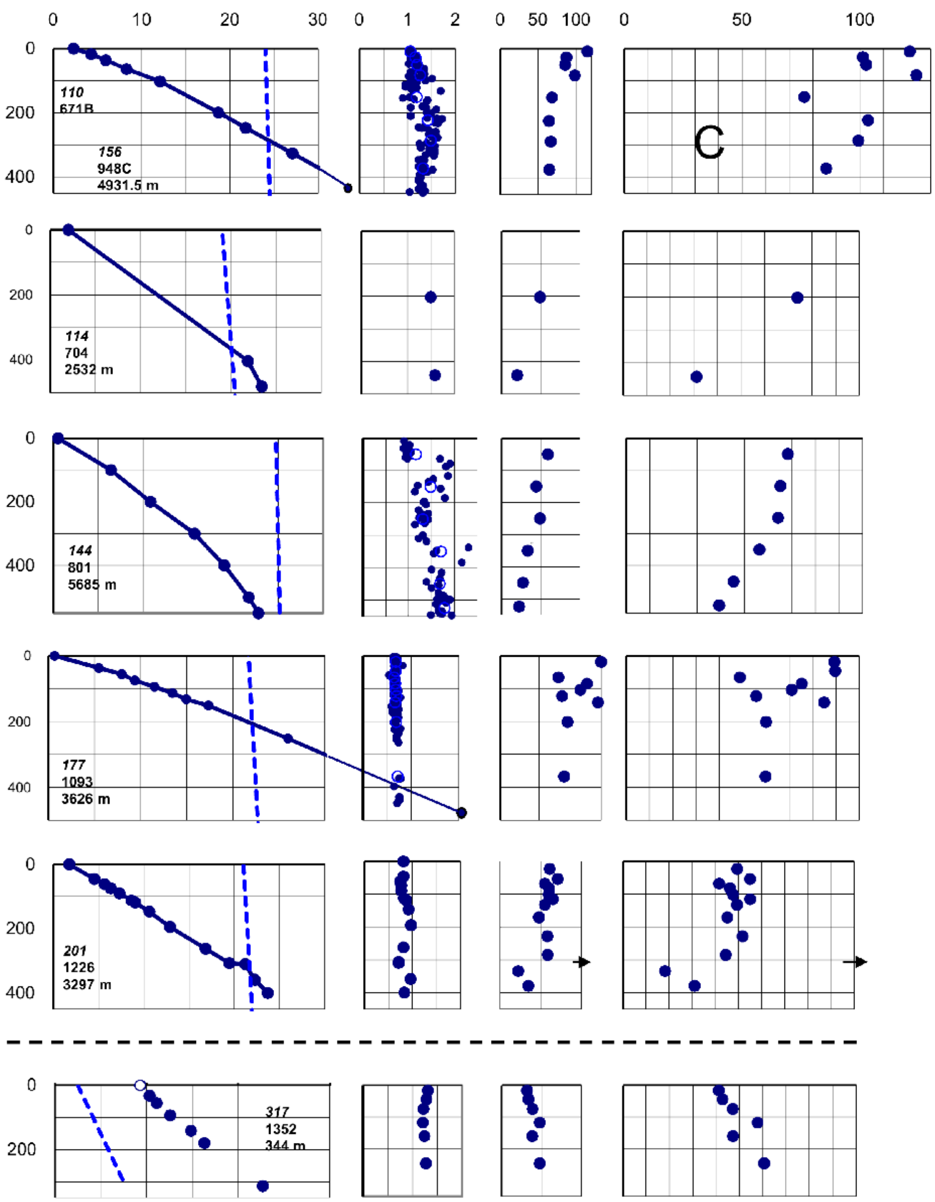

Temperature
°C
Thermal conductivity
W/m °C
Gradient
°C/km
Heat flow
mW/m 2
0
10
20
30
0
1
2
0
50
100
0
50
100
0
200
400
600
19
184
1910 m
19
185
2110 m
A
21
206
3196 m
22
214
1655 m
22
217
3010 m
25
242
2275 m
26
253
1962 m
26
254
1253 m

1333 **Fig 2b**



1334

Fig 2c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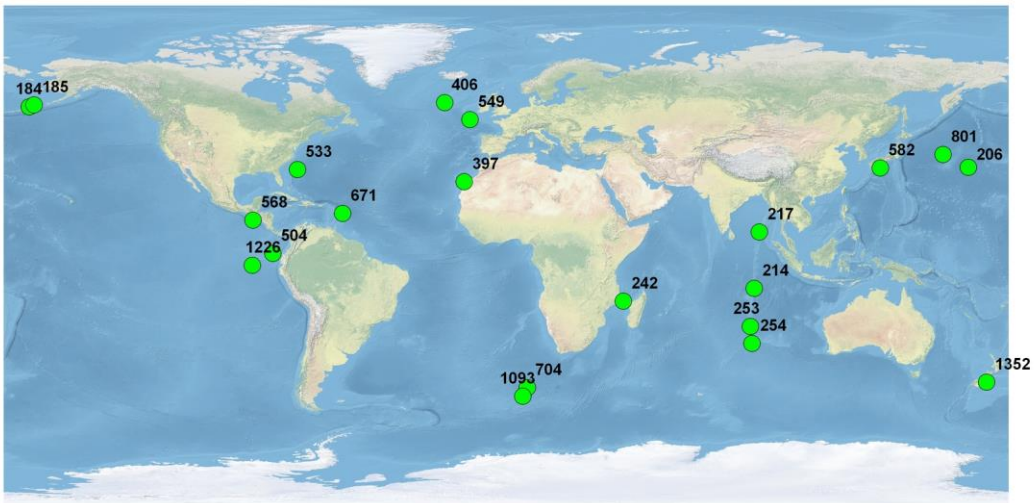
0 1 2
0 50 100
0 50 100
0
200
400
110
671B
156
948C
4931.5 m
C
114
704
2532 m
144
801
5685 m
177
1093
3626 m
201
1226
3297 m
317
1352
344 m

Fig 3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1093
1352

Fig 4

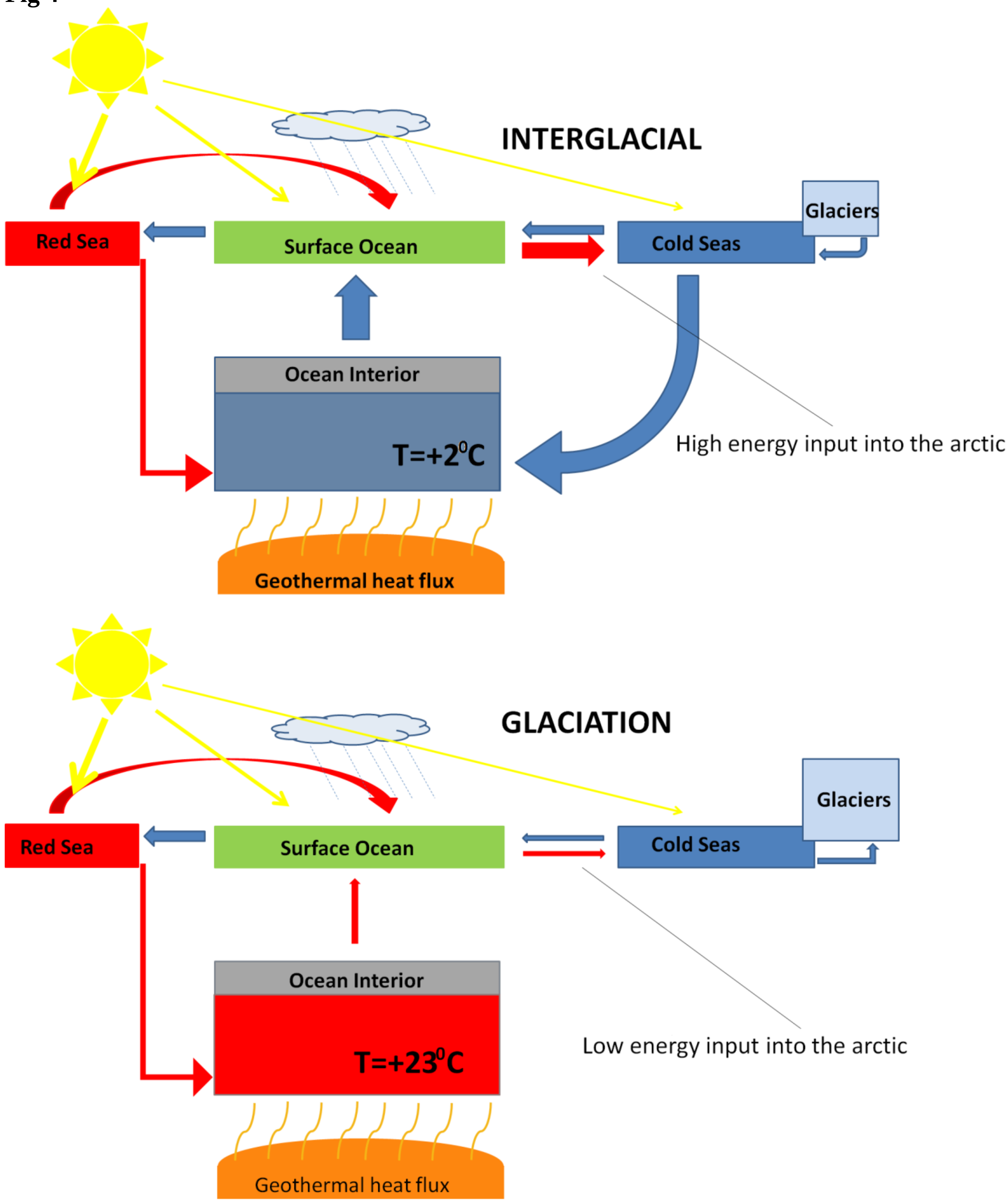
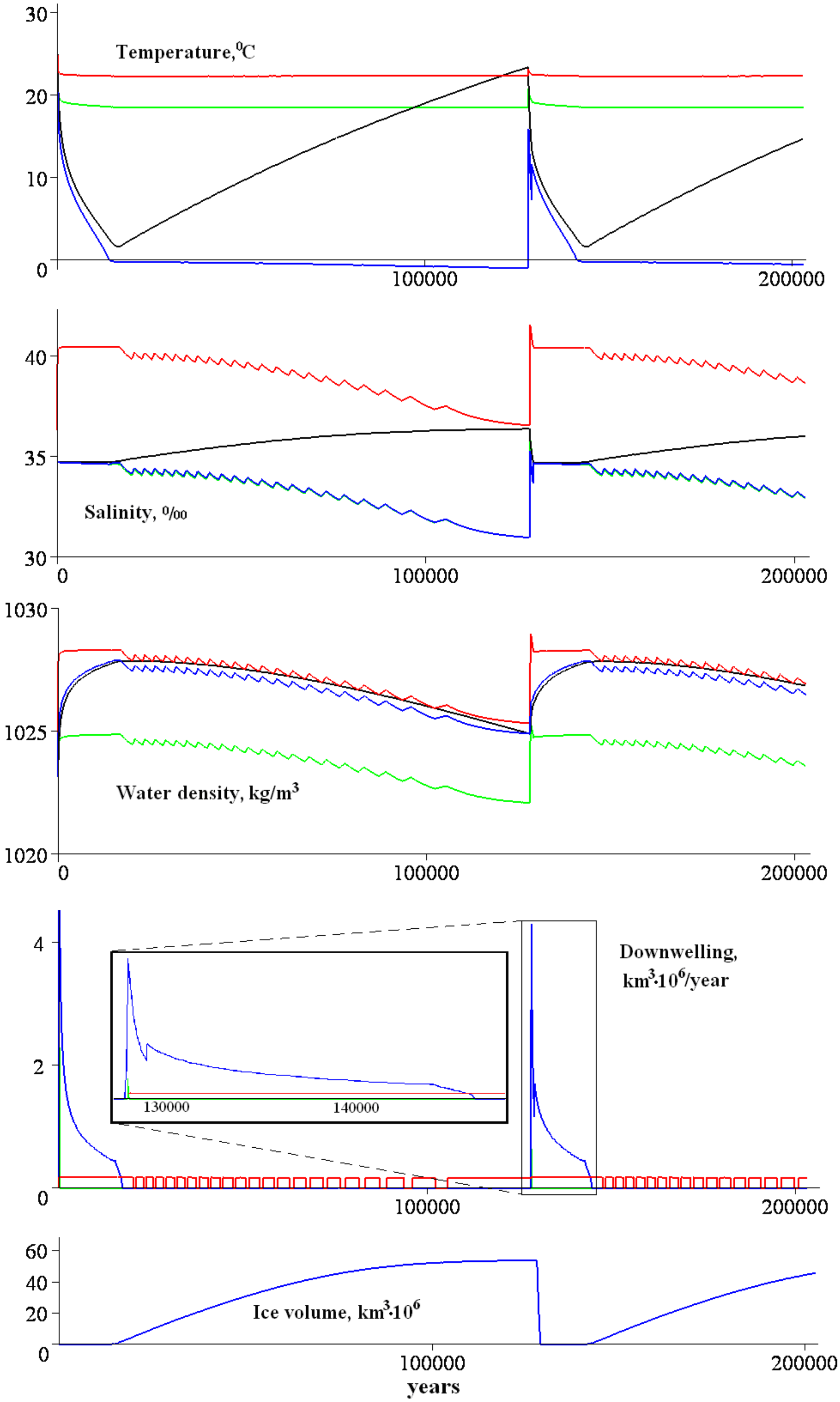

INTERGLACIAL
Red Sea
Surface Ocean
Cold Seas
Glaciers
Ocean Interior
T=+2⁰C
High energy input into the arctic
Geothermal heat flux
GLACIATION
Red Sea
Surface Ocean
Cold Seas
Glaciers
Ocean Interior
T=+23⁰C
Low energy input into the arctic
Geothermal heat flux

Fig 5


Temperature,⁰C
0
10
20
30
100000
200000
Salinity, ⁰/₀₀
30
35
40
0
100000
200000
Water density, kg/m³
1020
1025
1030
0
100000
200000
Downwelling,
km³·10⁶/year
0
2
4
130000
140000
100000
200000
Ice volume, km³·10⁶
0
20
40
60
100000
200000
years