A new earth's climate system model of intermediate complexity, PlaSim-ICMMG-1.0: description and performance

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Abstract. We describe a new intermediate complexity model (PlaSim-ESM-ICMMG-v.1.0) constructed by coupling the Planet Simulator [1-2] with ocean and sea-ice models. We demonstrate the results of climate simulation using PlaSim-ICMMG-v.1.0 considering global fields of surface air temperature, precipitation, sea surface temperature, and ocean circulation, and make comparisons with the results obtained using the original PlaSim version. PlaSim-ICMMG-v.1.0 reproduces the main features of the climate system reasonably well and demonstrates that it is useful for climate system modeling. Due to the rapid warming in the Arctic, there are major challenges associated with the mechanisms that regulate the dynamics of weather in the mid-latitudes. The new earth's system model of intermediate complexity PlaSim-ICMMG-v.1.0 can be used to deal with these challenges.

1.Introduction

Intermediate-complexity models are models describing the dynamics of the atmosphere and/or ocean in less detail than conventional General Circulation Models (GCMs). At the same time, they go beyond the approach taken by atmospheric Energy Balance Models (EBMs) or ocean box models by using sophisticated parameterizations of the unresolved flow or by explicitly resolving the equations of geophysical fluid dynamics, although at a coarser spatial resolution. Being computationally fast, intermediate-complexity models are capable of treating slow climate variations. Hence, they often include components of the climate system that are associated with long-term feedbacks like ice sheets, vegetation, and biogeochemical cycles. There exist many various approaches to building such a reduced model, resulting in a 'spectrum of Earth's system Models of Intermediate Complexity closing the gap between EBMs and complex GCMs [1-10].

Over the last few decades, the Arctic region has been warming much more than in the global average, primarily in winter, while the Arctic sea-ice extent has been decreasing dramatically [11-14]. Many mechanisms are at work in the 'Arctic amplification'. The existing positive snow and sea-ice albedo feedback plays a significant role in amplifying the warming signal [15]. The albedo feedback operates in summer when solar radiation is maximal. When the sea ice is shrinking and the water surface is opening, the warming due to the absorbed shortwave radiation can be large and enhances the sea-ice loss through lateral melting. Beyond the atmospheric heat transports, the high-latitude response to the greenhouse forcing involves an anomalous ocean heat transport into the Arctic. This occurs even if a weakened meridional overturning circulation (MOC) diminishes heat transport in the lower

latitudes. In addition, the ocean can act as a reservoir for the heat gained in summer when the sea ice retreats, storing it in winter months when the sea ice comes back.

The sea ice reduction is accelerating in the Barents and Kara Seas. The downward longwave radiation is an essential element of the sea ice reduction, but can only be sustained by an excessive upward heat flux from the sea surface exposed to air in the region of sea ice loss. The increased turbulent heat flux is used to increase the air temperature and specific humidity in the lower troposphere, which, in turn, increases the downward longwave radiation. This feedback process is clearly observed in the Barents and Kara Seas in the reanalysis data. A quantitative assessment reveals that this feedback process is amplifying at a rate of ~8.9 % every year in 1979-2016. Based on this estimate, the sea ice will completely disappear in the Barents and Kara Seas by around 2025. The excessive heat flux is necessary for the maintenance of this feedback process; a similar mechanism of sea ice loss is expected to take place in the sea-ice covered polar region when the sea ice is not fully recovered in winter.

In addition to these processes, the warmed ocean mixed layer delays the sea-ice growth, and, thus, affects the wintertime surface temperatures through a thinner ice pack. Since the Arctic atmosphere is stably stratified by thermal inversion at the surface, any warming that occurs there does not go far up into the troposphere.

There is a challenge whether the Arctic amplification affects the weather events in the midlatitudes [15-22]. What are the mechanisms? The rapid Arctic warming is expected to affect the weather patterns in the northern hemisphere, but how, where, and when is the topic of a study. Extreme weather events have been observed more frequently in the recent decades. A new earth's system model of intermediate complexity (PlaSim-ICMMG-v.1.0) has been constructed to study these problems as well.

The Integrated Earth's system model PlaSim-ESM-ICMMG-v.1.0 has been developed as a modular framework allowing a spectrum of intermediate-complexity Earth's system models to be created by selecting different options for the various climate and carbon cycle components. The plaSim-ESM-INMMG-1.0 model is capable of integration over multi timescales long runs. The framework has been designed to be modular to facilitate the coupling of more complex component modules as the available computing power increases.

2. Description of Earth's Climate System Model Plasim-ECS-ICMMG-1.0

2.1. PlaSim Version

PLASIM is a climate system model of intermediate complexity [1].

The PlaSim atmospheric dynamics is solved using a spectral transform method formulated for temperature, specific humidity, log surface pressure, divergence, and vorticity, and is described in detail in [2]. The shortwave radiation scheme used separates solar radiation into two bands, $\lambda < 0.75$ μm (with cloud scattering, ozone absorption, and Rayleigh scattering) and $\lambda > 0.75$ μm (with cloud scattering, cloud absorption, and water vapor absorption). The longwave radiation scheme uses a broadband emissivity method, with the (greenhouse gas) effects of water vapor, carbon dioxide, and ozone included in the calculation of emissivity. The ozone concentration is prescribed with an analytic spatio-temporal distribution. The cloud emissivity is calculated from the cloud liquid water content. The fractional cloud cover is diagnosed from the relative humidity (stratiform clouds) and from the convective precipitation rate (convective clouds). The other parameterized processes are largescale precipitation, moist convection (both cumulus and shallow), dry convection, boundary layer heat fluxes, vertical diffusion, and horizontal diffusion. A simple terrestrial dynamic global vegetation model land surface scheme, SimBA [2] is used to obtain the land surface variables.

The local runoff is transported to the ocean by a river transport scheme with linear advection.

The ocean and sea ice modules provide sea surface temperatures T_{sea}, sea ice distributions c_{ice}, and surface temperatures over the sea ice T_i. The ocean mixed layer heat flux (Qo) represents the oceanic transport and the deep water. The slab ocean model consists of a prognostic equation at each ocean

point for the oceanic mixed-layer temperature T_{mix}. [2] (Planet Simulator -Reference Manual Version 15 by F. Lunkeit et al., 2007).

2.2. PlaSim-ESM-ICMMG-1 Version

2.2.1. Ocean and Ice Models. As a basis for describing the oceanic block of the climate system, a model of the Arctic and North Atlantic developed at the Institute of Computational Mathematics and Mathematical Geophysics SB RAS is taken. This model was developed as a result of a significant improvement [24] in an earlier model of the World Ocean [25, 26]. The model is based on the traditional equations of ocean thermodynamics in curvilinear orthogonal coordinates using the hydrostatic and Boussinesq approximations. To specify surface boundary conditions, the "rigid lid" approximation is used. After vertical integration of the equations of fluid motion, the solution reduces to finding an integral component (the barotropic mode) and a deviation from it (the baroclinic mode). For the barotropic mode, the equations of motion are formulated in terms of an integral stream function. When integrating over time, a hybrid explicit-implicit scheme is used, splitting into physical processes and spatial coordinates. A more detailed numerical solution is described in [27]. In this version of the numerical model, the advective-diffusion operator is split into two operators describing the advection and diffusion process separately. To approximate the horizontal and vertical advection, a numerical scheme, ULTIMATE QUICKEST, proposed in [28] is used. The parametrization of vertical turbulence is based on the Richardson criterion and the vertical mixing procedure with the formation of a layer of a vertically homogeneous distribution of hydrodynamic characteristics. More details of the procedure can be found in [26]. Restoring of the surface salinity is not used.

The integration of the ocean module is carried out in conjunction with a model of thermodynamics of elastic viscous-plastic ice, CICE 3.1 [29], which is a modification of an earlier viscous-plastic Hibler's model [30].

As external characteristics determining the conditions for the development of sea ice, the following is used: the wind velocity, ocean surface velocity, specific humidity, density and temperature of the surface air, incoming short-wave and long-wave radiation, precipitation rate in the form of rain and snow, temperature and salinity of the ocean surface, and ocean surface tilt. The ice model calculates the following characteristics at the ice-water interface: the penetrating short-wave radiation, heat flux, salt and fresh water, and the frictional stress between water and ice.

In an earlier stand-alone version of the ice-ocean model, the characteristics of the lower atmosphere necessary for calculating the heat fluxes, moisture, wind friction stresses, and the downstream radiation fluxes were taken from the available data, for example, from the NCEP/NCAR reanalysis data. Within the climatic model system, the supplier of these data will be an atmospheric module. The initial distribution of the temperature and salinity fields corresponds to the climatic data of the Polar science center Hydrographic Climatology (PHC) [31] for winter. The initial ice characteristics were obtained as a result of a preliminary 10-year calculation with NCEP/NCAR 1948-1957 forcing.

2.2.2. Model Grid. The numerical grid of the model from 90S to 65N coincides with the lines of parallels and meridians with a resolution of 1×1º. At a latitude of 65N, this grid goes into a grid for the polar region constructed as a result of rotating a spherical coordinate system and re-projecting its hemisphere to a region above 65N (see [32] for more details). The horizontal resolution of the polar grid varies from 14 to 55 km with an average grid step of about 20 km. In the vertical direction, the grid consists of 38 levels with a maximum resolution of 5 m in the upper 20-meter layer. The minimum depth on the shelf is 20 m. The position of the shoreline is as close as possible to the geographic position taking into account the horizontal resolution. At the grid nodes where the depth was less than 20 m, the depth was artificially assumed to be 20 m.

The model domain includes the entire World Ocean. Some of the straits were artificially expanded in order to correspond to the model resolution.

2.2.3. River runoff. The river flow is calculated independently by a land model. Numerically, the river runoff is accounted for by additional precipitation, which is set at the grid points close to the geographic location of the river mouths.

2.2.4. Solid boundaries. The boundary conditions at the bottom and at the solid lateral boundaries ensure the absence of heat flows and salt flows through these boundaries. The bottom friction is proportional to the square of the near-bottom flow velocity; a no-slip condition is set on the lateral boundaries.

2.2.5. Parallelization. The modern problem of climate and climate systems research is increasingly associated with numerical modeling using high spatial resolution of the physical processes. The possibilities of solving it within the framework of serial numerical modeling have largely been exhausted. The creation and development of parallel models is an extremely topical activity. However, splitting into separate parallel modules is not enough, and parallelization of each module is required. The model of ice thermodynamics and ice drift CICE is based on explicit numerical schemes and is already parallelized based on a domain decomposition. The ocean model described above originally used implicit numerical schemes, which causes some difficulties in parallelization. Therefore, it was decided to translate the algorithms to use only explicit schemes horizontally and implicit ones vertically. This approach was implemented and has demonstrated its effectiveness in a number of test experiments.

3. Interaction between the climate system components

The model of the climatic system of the Arctic Ocean developed at ICMMG is a set of parallelly working and interacting modules responsible for the simulation of various components of the climate system. Until recently these were: an ocean circulation module, an ice thermodynamics module, and a drift module. The atmospheric module was organized in the form of a data model, i.e. in the form of a

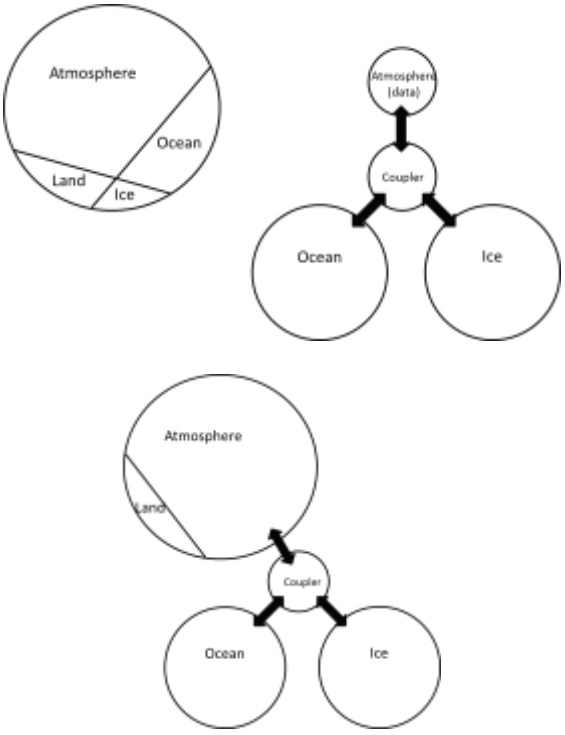


Figure 1. Graphical representation of the climate system components in the original version (a) and in the final variant (b).

set of programs for reading reanalysis data and their subsequent interpolation to the model grid. In addition, the software package includes a separate module called "coupler" which provides synchronization and interaction of the components (Figure 1a).

On the other hand, the PlaSim model already contains ocean and ice modules, whose quality does not fully correspond to the complexity of the tasks being solved and requires the replacement by modules of greater physical validity.

However, it is not just a question of replacing one set of subprograms by another set. In PlaSim, a grid that is common for all components is used, on the basis of which parallelism is performed by the method of domain decomposition. The use of a common grid is not advisable. Therefore, a variant was implemented where, instead of accessing the existing ocean and ice modules, a coupler module was called with necessary data exchange to form the flux fields between the system components (Figure 1b).

3.1. Coupler

In the original version, a coupler module was implemented in such a way that all components were supposed to work on the same grid. However, this approach, although it allows saving the calculation time due to the lack of interpolations, is not convenient. The grid used in the atmospheric module must be in a spherical coordinate system because of the need for spectral decomposition of the solution. The use of such a grid in the ocean model is inexpedient, because no spectral expansion is used here, and a direct numerical solution will be difficult due to a singularity of the coordinate system in the north pole region. Therefore, the ocean model grid should remain tripolar (such as described above). Moreover, this grid should allow an increase in the horizontal resolution, in contrast to the grid for the ice model, where the assumption of a statistical distribution of ice (used in CICE 3.1) is valid only with a rough resolution. In addition, the World Ocean contains areas where ice is never formed, and it would be optimal to exclude them from the sea-ice model domain. Therefore, a new revision of the coupler was formed based on the assumption that each of the interacting modules has its own grid and solution area, and it is required to provide not only their time synchronization, but also rapid interpolation from one grid to another. The numerical integration of the entire system in time should be organized so that the slowest modules work smoothly, i.e. the waiting time for a response from the coupler should be minimized.

3.2. Data exchange

The interaction of the modules is carried out by means of the coupler, which receives and sends the necessary data between the system components. Below is a table of data obtained by the coupler from each of the interacting modules, indicating in which modules this data is used:

Data short name	Long name	Used in
Atmosphere and Land Model		
zlvl	Wind velocity height	Ocean, Ice
ztlvl	Temperature and humidity height	Ocean, Ice
uatm, vatm	Wind velocity components	Ocean, Ice
potT	Potential temperature	Ocean, Ice
Tair	Air temperature	Ocean, Ice
Qa	Specific humidity	Ocean, Ice
rhoa	Air density	Ocean, Ice
Fsw	Downward shortwave radiation	Ocean, Ice
Flw	Downward longwave radiation	Ocean, Ice
Frain	Precipitation rate (rain)	Ocean
Fsnow	Precipitation rate (snow)	Ocean, Ice

cld	Cloudiness	Ocean, Ice
runoff	River discharge transport	Ocean
Ocean Model		
sst	Sea surface temperature	Atmosphere, Ice
sss	Sea surface salinity	Ice
uocn, vocn	Components of ocean surface current velocity	Ice
tiltx, tilty	Sea surface tilt components	Ice
frzmlt	Freezing potential	Ice
Ice Model		
ifrc	Ice area fraction	Ocean
tsfc	Temperature of the ice surface	Atmosphere
alb	Ice albedo	Atmosphere
tauxa, tauya	Wind stress	Atmosphere
lat	Latent heat flux	Atmosphere
sens	Sensible heat flux	Atmosphere
lwup	Upward longwave radiation	Atmosphere
evap	Evaporation rate	Atmosphere
Tref	Reference temperature at 10 m	Diagnostics
netsw	Net shortwave radiation	Atmosphere
melth	Heat flux into the ocean due to ice melt	Ocean
meltw	Fresh water flux into the ocean due to ice melt	Ocean
salt	Salinity flux into the ocean due to ice melt	Ocean
tauxo, tauyo	Ice-ocean stress	Ocean
swabs	Shortwave radiation passing through ice into ocean	Ocean

As a result of the coupler, the flow fields and their characteristics for different modules of the system are formed. Below is the distribution table for each module:

Data short name	Long name
Atmosphere and Land Model	
tsurf	Surface temperature
Cu	Drag-coefficient for wind stress ($\boldsymbol{\tau} = \rho C_u \mathbf{u}$)
Ct	Drag-coefficient for sensible heat flux ($H_s = \rho c_p C_T (T-T_s)$)
Ce	Drag-coefficient for evaporation flux ($E_s = \rho C_e (q-q_s)$)
albedo	Surface albedo
Ocean Model	
swave	Shortwave radiation
netheat	Sensible heat flux
prec	Precipitation rate
evap	Evaporation rate
taux, tauy	Sea-surface stress
ifrc	Ice area fraction
melt	Heat flux due to ice melt
salt	Salt flux due to ice melt
utidal, vtidal	Tidal velocity components

Ice Model	
sst	Sea surface temperature
sss	Sea surface salinity
uocn, vocn	Components of ocean surface current velocity
zlvl	Height of lower atmosphere characteristics
uatm, vatm	Wind velocity components
potT	Potential temperature
Tair	Air temperature
Qa	Specific humidity
rhoa	Air density
tiltx, tilty	Sea surface tilt
fw	Freezing potential
fsw	Downward shortwave radiation
flw	Downward longwave radiation
rain	Precipitation rate (rain)
snow	Precipitation rate (snow)

4. Results. 100-year coupled run

As a test, the model of the climate system was calculated for a period of 100 years. The initial state of the atmosphere was obtained in previous experiments with the full version of the stand-alone PlaSim model.

4.1. Ocean

Figure 2 shows the sea surface temperature obtained as a result of averaging over the last ten years of an experiment for March and September. In general, the temperature distribution is in good agreement with the known climate distributions, but it is about 2-3 degrees colder in the tropics. The results for the most critical ocean region, the Arctic, are not entirely satisfactory. The flow in the upper 100 m layer in Figure 3 corresponds to a strong cyclonic circulation mode and does not contain characteristic features of the anticyclonic circulation in the Beaufort Sea. As a result of insufficient convergence of the currents in this region, the ice field also turned out to be more dispersed than the real distribution, which led, for example, to the fact that an area north of Scandinavia and the Murmansk port were under ice (Figures 4a and b), even in summer. In summer (southern hemisphere), the ice field in the Antarctic (Figure 4c) is almost completely absent, and in winter it reaches 0.5-1 m (Figure 4d). The heat transport from the equator to the poles is somewhat overestimated in the northern hemisphere and underestimated in the southern hemisphere. However, if we consider the meridional heat transport in dynamics, it can be seen that a 100-year period is probably insufficient for its final formation, and the changes that have occurred in the last 50 years are in a positive direction (Figure 5).

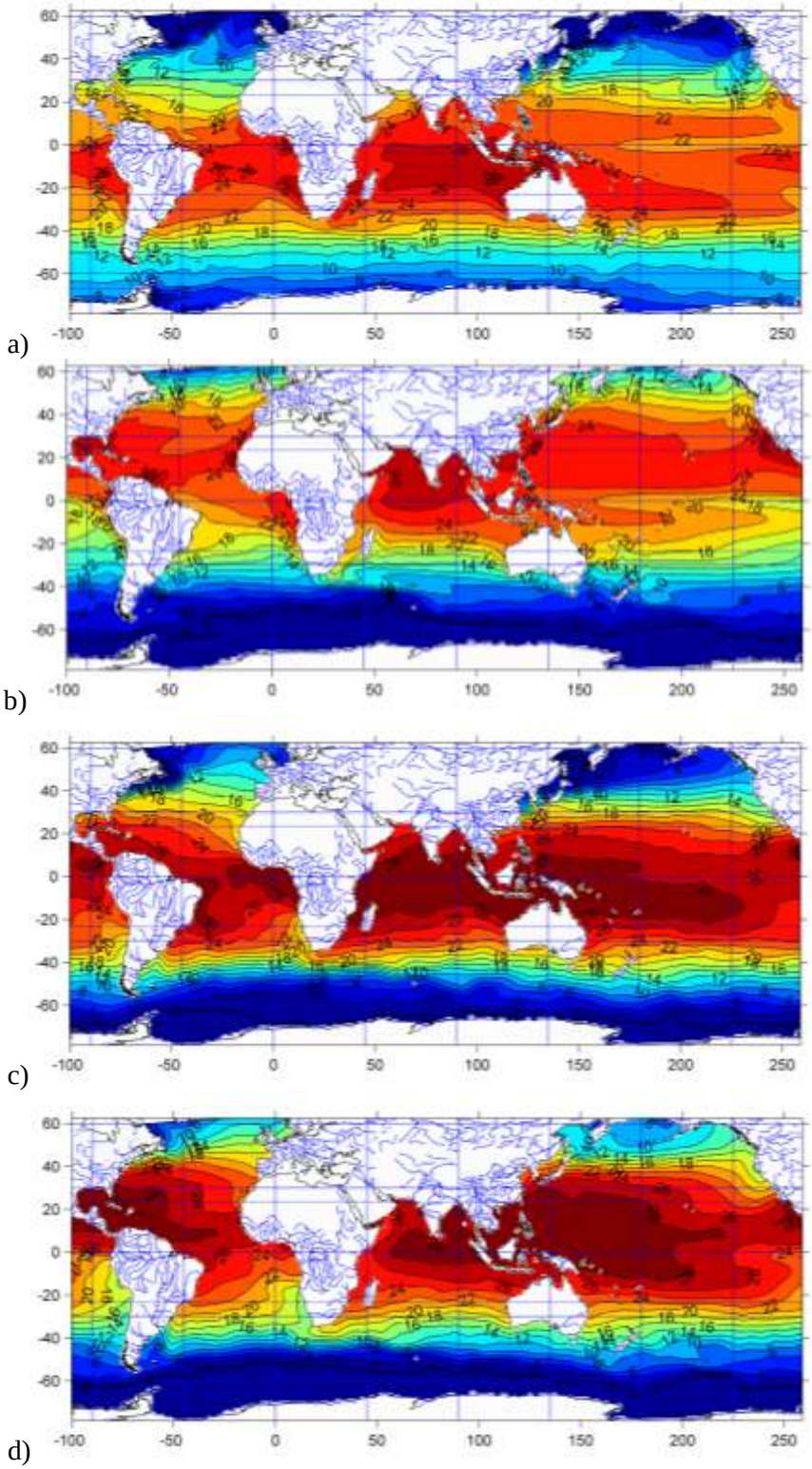
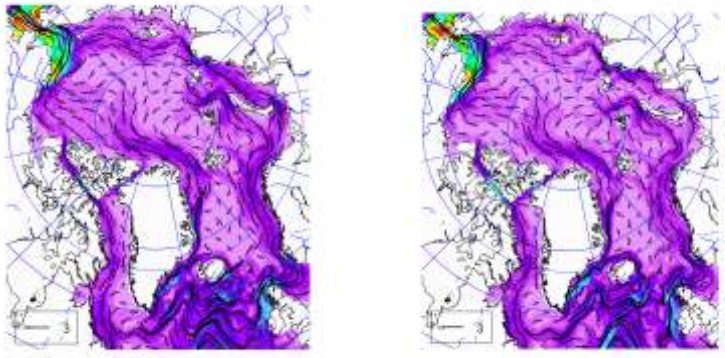
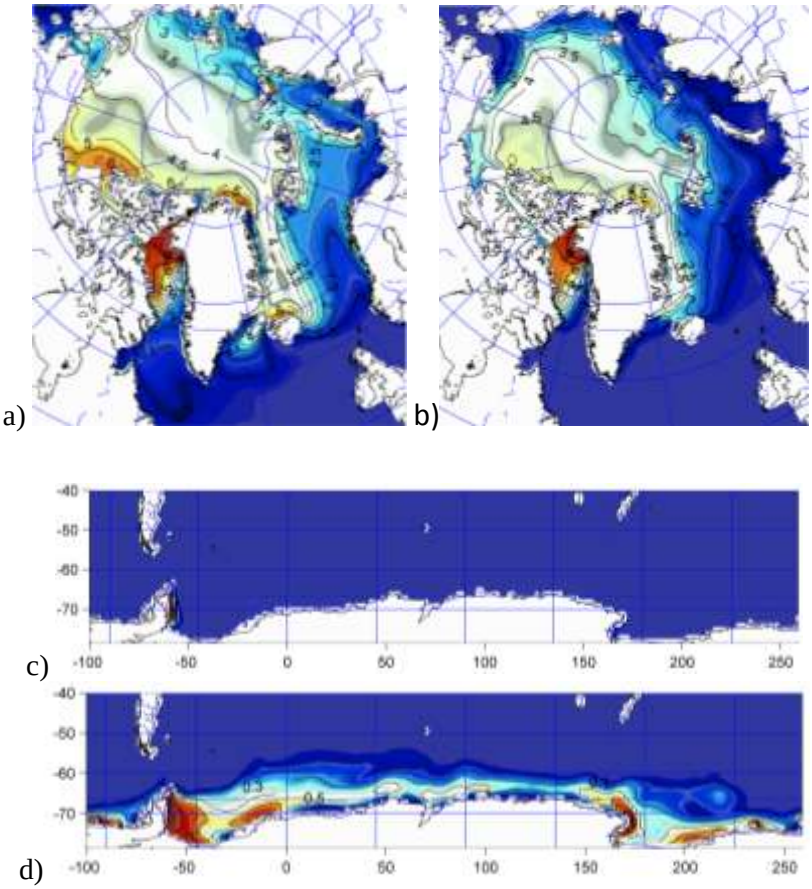


Figure 2. The sea surface temperature a) in March b) in September, averaged over the last decade of the 100-year preliminary experiment. The PHC3.0 climatological sea surface temperature c) in March d) in September [Steele et al, 2000].



a) b)

Figure 3. The average circulation (cm/s) of the upper 100-meter layer of the ocean in the Arctic: (a) - in March, (b) - in September.



a) b)

c)

d)

Figure 4. Ice thickness (m) in the Arctic: (a) in March, (b) in September and in the Antarctic: (c) in March, (d) in September.

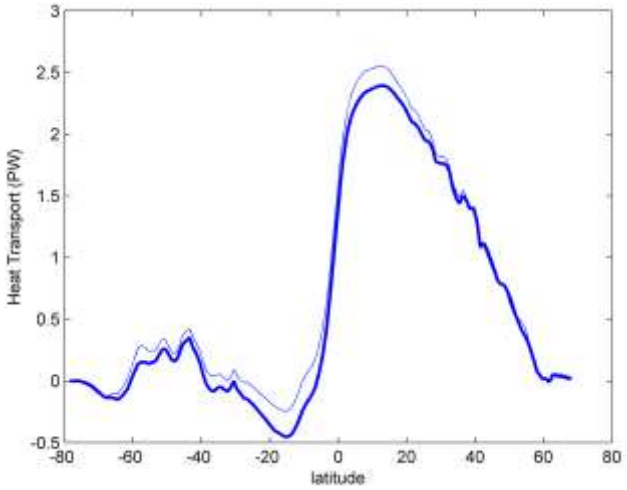


Figure 5. Meridional heat transport (PW – 10^{15} W, a positive value corresponds to a northward transport). The thin line corresponds to the averaging over the year-period 45-55, and the bold line -- over the year-period 90-100 of the preliminary 100-year experiment.

4.2. Atmosphere

Surface air temperature and precipitation fields are plotted in Figures 6-9. Figure 6 shows the surface temperature obtained as a result of averaging over the last 35 years of the experiment and over the 1979-1999 of the NCEP2 Reanalysis data for January and July.

In general, the model surface temperature distribution is in good agreement with the NCEP2 Reanalysis data, but some regions are overheated. It is about 5 degree warmer in the central part of Eurasia in January, in the southern part of this region in July, in the western part of South America in January and in North America in both months considered. In the Polar regions the model surface temperature is lower than that in the Reanalysis data in its winter seasons (in January for the North Pole and in July for the South Pole). Figure 7 shows the total precipitation obtained as a result of averaging over the last 35 years of the experiment and over the 1979-1999 of the GPCP ver. 2.3 data (www.esrl.noaa.gov) for January and July. The model data shows a good agreement with the observation data, but it is about 10-15 mm per day greater in the modeling data than in the observations in the tropics for both months considered. Some major arid regions are captured, specifically the seasonal migration of the Inter-Tropical Convergence Zone and the associated monsoon systems. The PlaSim-ICMMG-v.1.0 better correlates with the observed seasonal precipitation than the PlaSim, and it shows wetter conditions (Figure 8).

The results of modeling with the new model show that in January in the Arctic the global surface temperature is lower than the observed one: a cold bias and a comparison of the simulation results with the old model with the reanalysis data shows a warm bias (Figure 9).

The average observed January and July climatology is better simulated by the PlaSim -_ICMMG-v.1.0 model than by the PlaSim model.

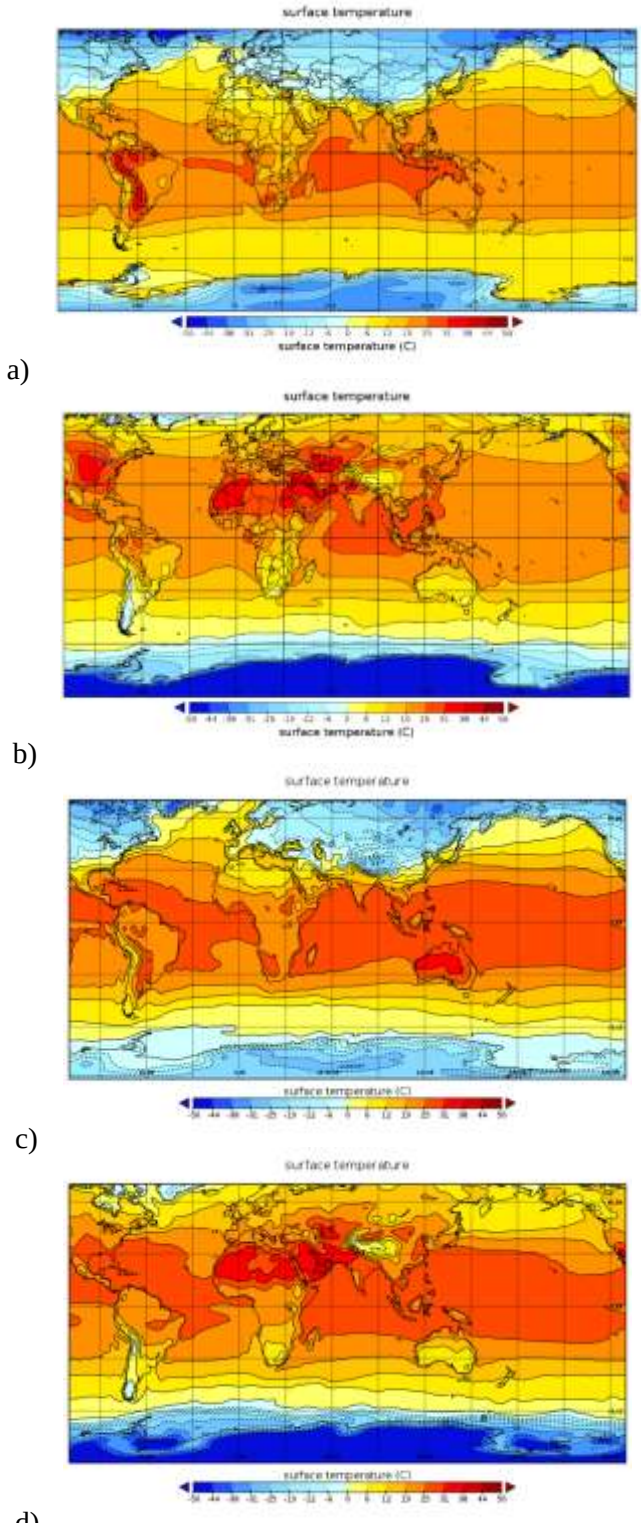
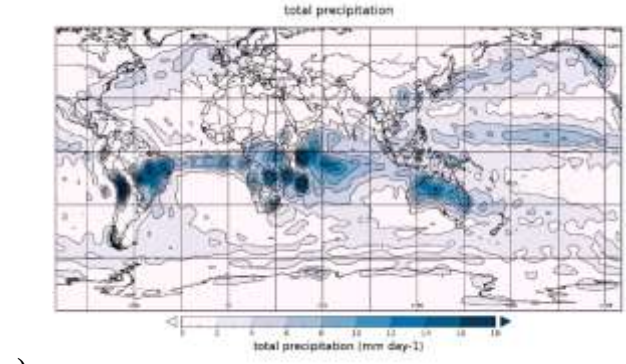
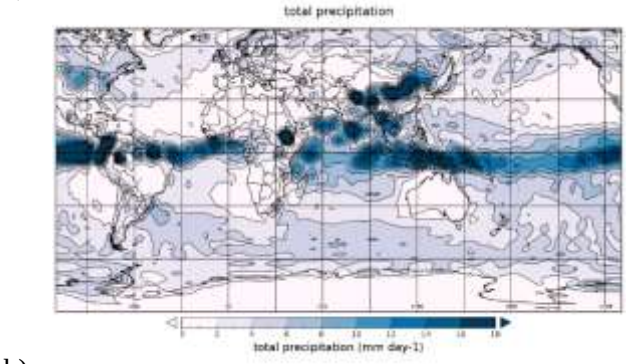


Figure 6. Surface temperature averaged over the last decade of the 100-year preliminary experiment: a) in January; b) in July. The climatological surface temperature for NCEP2 Reanalysis averaged over 1979-1999 for c) January and d) July.



a)



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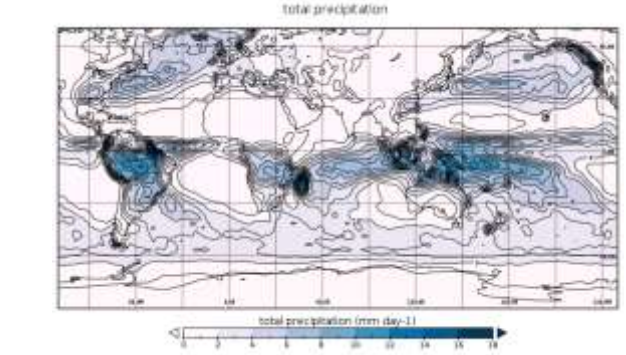

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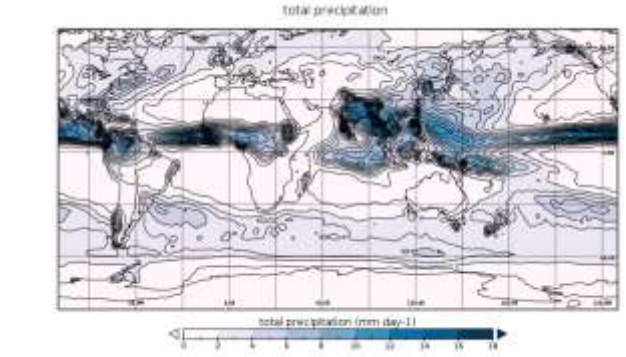

d)

Figure 7. Combined precipitation averaged over the last decade of the 100-year preliminary experiment: a) in January; b) in July. The climatological total precipitation for GPCP ver. 2.3 dataset averaged over 1979-1999 for c) January and d) July (www.esrl.noaa.gov).

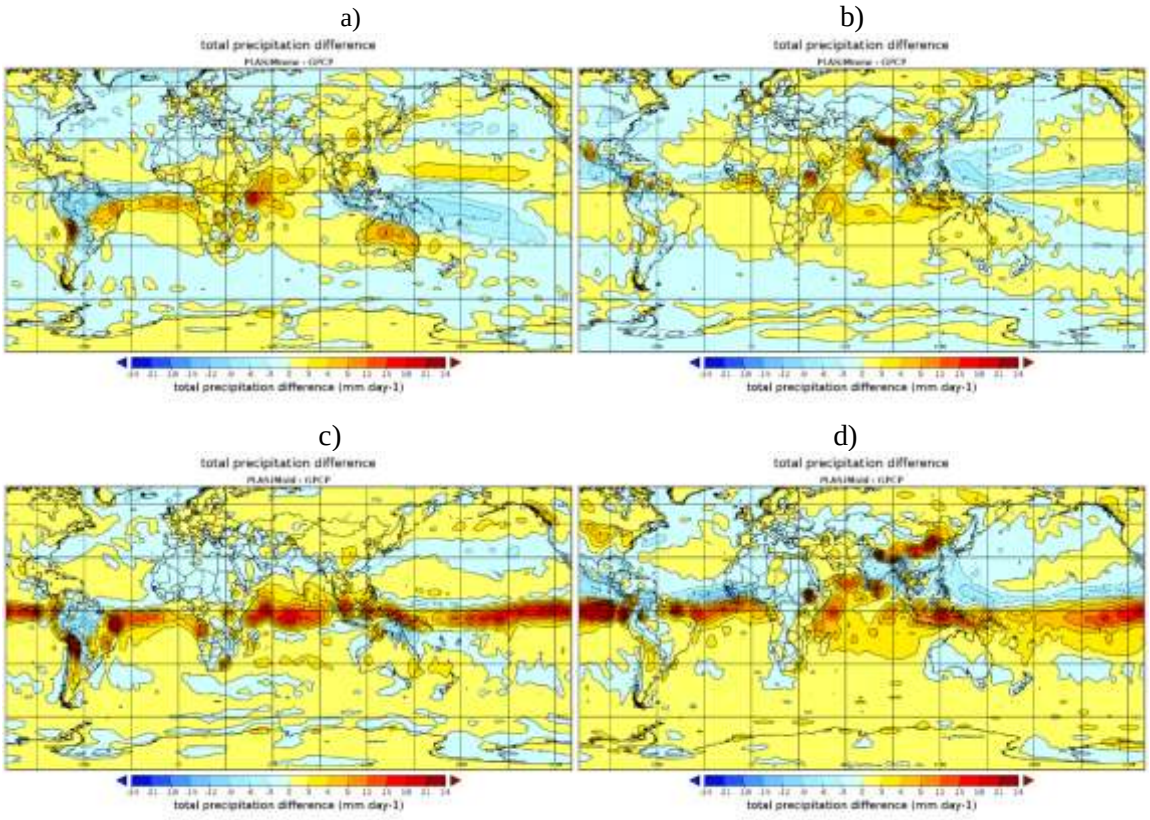

Figure 8. Differences between combined precipitation averaged over the last decade of the 100-year preliminary experiment and climatological total precipitation for GPCP ver. 2.3 dataset averaged over 1979-1999(www.esrl.noaa.gov): a) in January; b) in July for PlaSim-ICMMG-v.1.0 ; c) January and d) July for PlaSim.

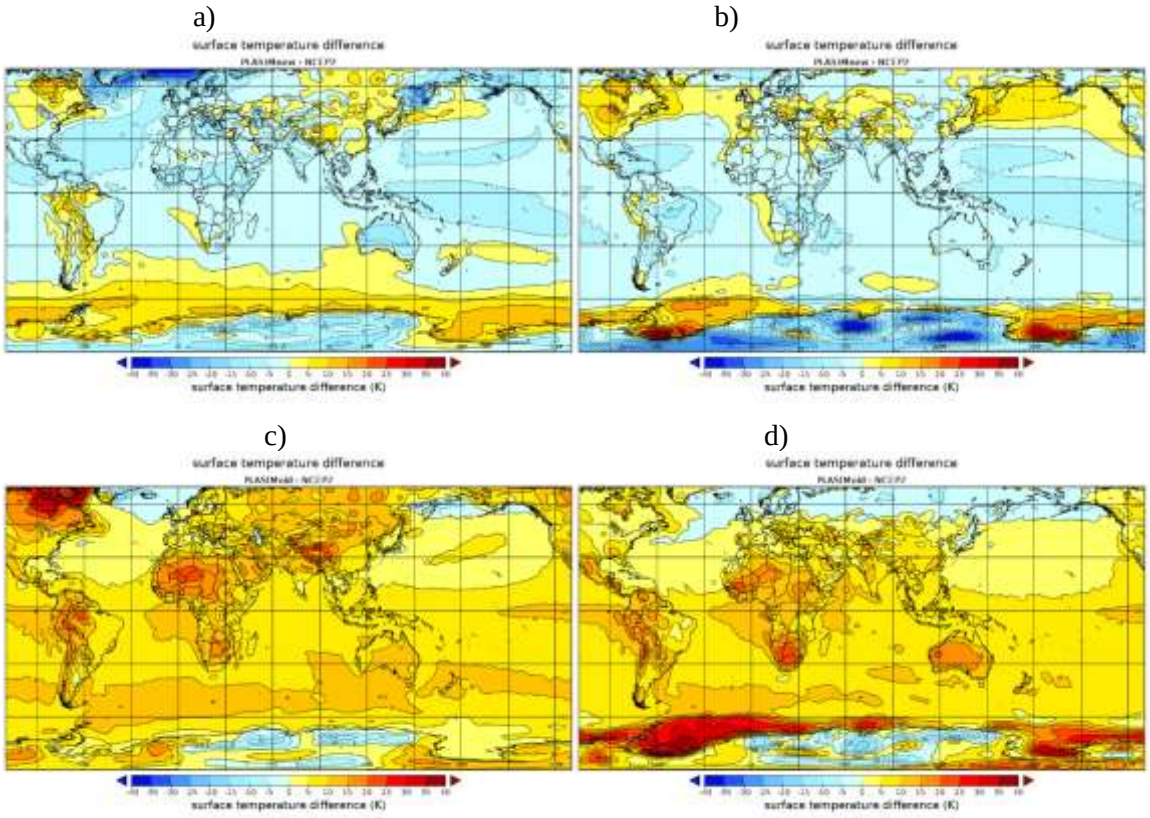

Figure 9. Seasonal surface air temperature (C) difference: (PLASIM-ICMMGv.1.0 - NCEP) - a) January, b) July; (PLASIM - NCEP) – c) January, d) July.

5. Conclusions

It is well-known that the complexity of the climate system presents a challenge to the climate theory. We try to simulate as much of climate dynamics as we can with comprehensive numerical models (for example, [33]), and we also try to understand the simplifying and highlighting key phenomena in idealized models [34].

We have presented a new intermediate complexity climate system model, PLASIM–INMMG-v.1.0, which reproduces the main parameters of the climate system. We conclude that the new model reproduces the seasonal surface temperature and precipitation reasonably well and in general better than the old version, PlaSim.

The features of the ocean circulation reproduced by the model and the resulting distribution of sea ice significantly deteriorated in comparison with the results obtained earlier [24]. The sea surface temperature in the equatorial region is 2-3 degrees below the climate data. The ice distribution is wider towards the mid-latitudes in the North Atlantic, while in the southern hemisphere the ice almost completely disappears in summer (in the southern hemisphere). Nevertheless, the dynamics of the formation of meridional heat transfer is such that it allows us to state that, firstly, a quasi-stationary state of the climate system has not yet been reached in the first 100 years of model integration. Secondly, from the point of view of heat exchange between the northern and southern hemispheres, the system develops in a positive direction, allowing us to hope that a quasi-stationary state will be achieved with climatic parameters closer to the observed ones. The achieved state, despite its difference from the observed climate, is still quite plausible, which means that the model needs some additional tuning and selection of more suitable values of the model parameters.

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