The *USSP2011* Hitchhikers Guide to the Black Arts
(of Earth system modelling)

VI: 'How low can you go?'

Stuff to keep in mind:

"There are known knowns.
These are things we know that we know.
There are known unknowns.
That is to say, there are things that we know we don't know.
But there are also unknown unknowns.
There are things we don't know we don't know."

Donald Rumsfeld, former US Secretary of Defense

12. The last glacial

12.1 You will be using a 'modern' configuration of *c*GENIE, but it is rather more idealized than you have seen for the modern Earth (although not quite as idealized as for the snowball Earth experiments). It also differs in that in addition to having a carbon cycle (which was omitted from the majority of your previous session) it includes a representation of deep-sea sediments and interaction between the preservation and burial of $CaCO_3$ and ocean chemistry. For an over-view of the sediment model and what time-scales and nature of carbon cycle interaction between ocean and sediment you can expect – read: *Ridgwell and Zeebe* [2005] (over-view of the role that calcium carbonate production by life and deep-sea sediment play in the global carbon cycle) and *Ridgwell and Hargreaves* [2007].

Note that you will be using a new *c*GENIE configuration, so before you forget ... `make cleanall` … You will also be using a different 'run genie' script: `old_rungenie_t48` when running the model (it configures different (faster) time-stepping).

12.2 Take the new model for a spin by running on from the *re-start*: `exp0_glacial_SPINUP`. This is a steady-state climate+carbon cycle experiment that includes the deposition of $CaCO_3$ in deep-sea sediments and the balance between weathering (solute input to the ocean) and burial (output). Try running ('briefly', but 100 years would not be too tedious for this faster configuration!):

```
./runcgenie.t48.sh cgenie_eb_go_gs_ac_bg_sg_rg.p0000b.BASESFe.t48
 LABS exp13_glacial_CONTROL 100 exp0_glacial_SPINUP
```

Note that the *base-config* (`cgenie_eb_go_gs_ac_bg_sg_rg.p0000b.BASESFe.t48`) is different from before (and specifies the use of a sediment model '`sg`' in GENIE for instance). It also includes an iron cycle alongside phosphate as a limiting nutrient (the '`Fe`' bit) and takes fewer time steps per year ('`t48`'). In fact, you'll notice that is it rather faster than previous configurations :)

The degraded resolution (and fewer time-steps per year) is important in being able to run *c*GENIE on sediment time-scales (see *Ridgwell and Hargreaves* [2007]) within a reasonable time-scale (i.e., the time available to you for carrying out some glacial CO_2 research using the model). Unfortunately, that the resolution is rather more degraded than previously means that you will need to be aware of additional limitations and caveats associated with this configuration than previously (these limitations and caveats are left for you to identify and take on board).

The *user-config* is set up as a 'control' in that it continues on from the *re-start* without making any adjustments to the climate or biogeochemsitry (compared to the *re-start*) but will show up any residual drift in the spin-up. Note that the *user-config* has been set up so that the global carbon cycle is 'open' – that is to say, that there is an input of carbon (and alkalinity) to the ocean from weathering, and a loss due to preservation and burial of $CaCO_3$ in deep-sea sediments. Depending on the state of ocean chemistry (and biology) and weathering, these two fluxes (input and output) do not have to balance, and hence ocean carbonate chemistry can change with time. The spin-up may not have the two fluxes (input and output) perfectly balanced, hence a '*control*' experiment will reveal any residual drift. A residual drift can be dealt with if it is relatively small and near linear and you have a *control*, because any experiment you carry out will likely also incorporate (or be biased) by the same residual drift, hence running a *control* gives you something to directly contrast with – your experiment minus the *control* (e.g., a difference map or simple subtraction of global numbers) will give you the effect of whatever parameters you changed in the experiment and corrected for any drift. (Note that in previous Labs we were a bit lazy, and difference maps were often created with respect to year 1 of an experiment – strictly, they should have been created relative to the same year of a parallel *control* experiment, i.e., results at year 100 should have been contrasted with the year 100 results of the *control*.) Depending on exactly how you design your experiments in investigating glacial CO_2; a single control experiment might suffice.

There is a new set of outputs from this configuration of *c*GENIE, including sediment output (from the 'SEDGEM' module). For instance, the composition of the sediments at the very end of a model experiment (hence unlike BIOGEM, which saves a series of time-slices) is saved by the SEDGEM module in the `sedgem` subdirectory of your results directory in a netCDF file called `fields_sedgem_2d.nc`. (Note that there is some duplication of results saving, because a series of

time-slices of sediment composition are also saved in the 2D biogem netCDF file `fields_biogem_2d.nc` alongside with sea-ice extent etc.).

For instance, the 2D distribution of wt% $CaCO_3$ – which is the weight fraction of calcium carbonate ($CaCO_3$) in the surface sediments of the deep ocean (i.e., how much plankton carbonate shell material is there compared to other stuff in the mud at the bottom of the ocean?) is saved under a variable called: `sed_CaCO3`. How much carbonate material there is tells you both something about how many carbonate shell secreting plankton were living at the ocean surface above and what is the chemistry of the deep ocean like that these tiny shells were preserved and did not dissolve. To gauge to what degree the faster configuration of GENIE might provide an adequate representation of the interaction between ocean chemistry and sediment composition (e.g., in $CaCO_3$ buffering of CO_2 release and 'carbonate compensation'), the output should be contrasted to observational-based maps as well as (higher resolution) model results (e.g., as presented by *Ridgwell and Hargreaves* [2007]). Appreciate that *c*GENIE does not reproduce reality ... particularly at this lower resolution, but does it get the broad patterns right (is it more right than wrong, or more wrong than right)? Do you think the model-data misfits might be important? (Note that there is a reconstruction of the glacial pattern of $CaCO_3$ in sediments that is available and may be of use in constraining your glacial CO_2 hypothesis, or at least testing it against data.)

12.3 In light of *Kohfeld and Ridgwell* [2009], other model outputs/fields you might also initially 'consider' (i.e., quickly view now and bear in mind for later) for which some data/observational constraints on your glacial CO_2 'solution' may be available, include (but not as an exhaustive list):

- The 2D `fields_biogem_2d.nc` field: `phys_seaice` ('sea-ice cover'), for which some glacial sea-ice limit information exists.
- The 2D `fields_biogem_2d.nc` field: `ocn_sur_temp` ('surface-water temp') (or view the surface ocean layer in the 3D file), for which 2 (one older, one newer) comprehensive datasets exists – it would be reasonable to question whether you achieve an adequate glacial (surface) climate state and if not, whether this impacts any bias (and in which direction) to your CO_2 solution.
- The 2D `fields_biogem_2d.nc` field: `sed_CaCO3` ('sediment core-top CaCO3') (also available from the `sedgem` model output), for which some glacial $CaCO_3$ distribution data/estimates exist.
- The 2D `fields_biogem_2d.nc` field: `phys_opsia` (' Atlantic streamfunction'). While poorly resolved in this model configuration, many (glacial) model studies report the circulation field and hence they provide a point of comparison for your GENIE-based research.
- The 2D `fields_biogem_2d.nc` field: `ocn_D_DIC_13C` (' planktic-benthic difference DIC_13C') and also the individual planktic (surface) and benthic (bottom) fields. A significant amount of $\delta^{13}C$ data exists in the literature for both glacial and interglacial states.
- The 2D `fields_biogem_2d.nc` field: `ocn_ben_O2` (' bottom-water O2') (and also horizontal slices in the 3D file). Ideally, no-where in the ocean should anoxia (no oxygen) occur. It certainly should not be widespread across one or more ocean basins if your glacial CO_2 solution is to get published in Nature ;)
- The 2D `fields_biogem_2d.nc` field: `ocn_ben_sal` ('bottom-water sal') (and also in the 3D file as horizontal slices), for which some estimates exists for a few places in the ocean.
- The 2D fields for surface and deep PO_4 (and also in the 3D file as horizontal slices) as some proxy evidence exists for changes in nutrient utilization.

Also refer to the 3D netCDF files and the time-series where helpful.

(Note here that the spin-up provided is 'modern' and hence glacial data cannot be directly contrasted – these are suggestions/guidance for a starting point for later analysis.)

12.4 Before you run any experiments, confirm whether the *spin-up* provided really is adequately 'spun-up'. Or if not: how much and quickly does it drift, and in what properties of the Earth system (carbon cycle) does it drift most in? While the exercises in this Lab can be perfectly adequately carried out with a small residual drift (if one exists), for your glacial CO_2 investigation proper, you might want to think about either creating a new, longer spin-up, or continuing the current spin-up,

e.g., by running the control out for ... well, you will have to judge for yourselves how long to run it for ...

12.5   In addition to the control *user-config*, you have been provided with two additional ones: `exp0_glacial_SPINUP` is the *user-config* used to create the spin-up (this was run for a total of 20000 years), and `exp12_glacial` is a template *user-config* file for you to base (if you want!) your glacial CO_2 investigations on (at least initially).

12.6   In your glacial CO_2 investigations, 4 separate initial modifications of the model to nudge it towards a glacial state are provided (e.g., see *Kohfeld and Ridgwell* [2009]).

- A modification of surface (actually 'planetary') albedo to try and take account of some of the cooling influences of the large (Northern Hemisphere) ice sheets that were present during the last glacial but which are not calculated or explicitly taken into account in the version of GENIE you are using.
- A modification of greenhouse gas radiative forcing (as per in the snowball Earth experiments) to take into account the lower CO_2, CH_4, and N_2O concentrations in the atmosphere during the last glacial. While you will be attempting to reproduce ~190 ppm atmospheric CO_2 (and hence deduce the reasons for low glacial CO_2), you may not necessarily achieve this, and you have no means of explicitly controlling the other greenhouse gases, so you may as well get the radiative forcing and hence the glacial climate state as close as possible before trying to adjust the carbon cycle. But it is up to you whether you prefer to not 'cheat' and have whatever CO_2 *c*GENIE simulates, directly affect climate (and hence feed back on CO_2).
- The approximately ~3% (~1 PSU) increase in salinity (and other dissolved tracers) due to the presence of large (Northern Hemisphere) ice sheets, and hence loss of freshwater from the ocean and lower sea-level.
- The release of carbon stored on land in vegetation and soils during glacial climates. *c*GENIE has some capabilities to model changes in terrestrial carbon storage, but you are not using a version with this science module enabled.

You will probably want to carry out separate experiments to test the effect of each of these in turn and hence to learn the effect and impact on atmospheric CO_2 of each individually before combining them. You can implement them as follows:

- For the pseudo-glacial planetary albedo modification, add the following lines to a *user config* file:
```
# adjusted planetary albedo
ea_albedop_offs=0.200
ea_albedop_amp=0.360
ea_albedop_skew=0.0
ea_albedop_skewp=4
ea_albedop_mod2=-15.000
ea_albedop_mod4=-2.500
ea_albedop_mod6=0.000
```
- For glacial radiative forcing:
```
# glacial CO2 radiative forcing
ea_radfor_scl_co2=0.6835
# glacial CH4 radiative forcing
ea_radfor_scl_ch4=0.5
# glacial N2O radiative forcing
ea_radfor_scl_n2o=0.8
```
- For a ~3% increase in salinity:
```
bg_ctrl_force_GOLDSTEInTS=.true.
bg_par_forcing_name='p0000b_FeMahowald2006_ADJUST_salinity'
```
- For the reduction in terrestrial carbon storage:
```
bg_par_forcing_name='p0000b_FeMahowald2006_ADJUST_terrestrialC'
```

Note that for the <u>first two</u>, you will need to include these lines in all (glacial) experiments that you require glacial albedo and radiative forcing for.

The situation for the <u>second two</u> is different – these are effectively one-off changes imposed on the global climate and carbon cycle. By selecting the salinity forcing, you add 1 PSU of salinity to the entire ocean (and concentrate proportionally all dissolved tracers in the ocean) in a single year. Obviously you only want to do this <u>once</u>, not multiple times (other wise you will get an increasingly salty ocean ...). You are arguably best off running a (further) spin-up to incorporate the salinity change, and then all experiment thereafter do not require a salinity change specified (because it has already been implemented). The advantage of this approach is that you keep separate the effects of the salinity change from other modifications you make to the system. Similar reasoning applies to the terrestrial carbon change. As specified, this forcing results in 500 PgC of carbon being added to the atmosphere over a period of 500 years (i.e., at a rate of 1 PgC yr^{-1}) as if to simulate a commensurate reduction in carbon stored on land. One strategy might be to implement this as a second phase of (additional) spin-up (after the salinity modification). Because, with the sediments, if takes some 10s of thousands of years for the system to fully-readjust, ultimately (not necessarily during the Lab) you might want to consider running both follow-on spin-ups for an extended period (to steady-state). Note that while the magnitude of the glacial-interglacial salinity and sea-level change is well constrained, that of the terrestrial biosphere is not (e.g., see: *Kohfeld and Ridgwell* [2009]). In investigating the potential causes of low glacial CO$_2$, do not feel constrained to necessarily run with the default (500 PgC) forcing provided ... (Think for yourselves!)

Play around with the climate boundary conditions (albedo and radiative forcing) and also try imposing the two geochemical boundary conditions to the ocean/atmosphere (salinity and terrestrial carbon). Questions you might want to ask may include: What effect do they have on atmospheric CO$_2$? What about ocean properties (for which proxy data exists) such as temperature, salinity, sea-ice extent, wt% CaCO$_3$, δ^{13}C, etc (e.g., see Section 11.3). What are the time-scales of response (there may be multiple time-scales and with CO$_2$ and other properties changing in different directions at different times)? Do the forcings combine linearly in their impact? Obviously, be asking these questions mindful of apparent model (and data) limitations.

13. Accounting for low glacial CO$_2$

A2.1 By now, you should have created a new spun-up model state, incorporating: (i) higher planetary albedo, (ii) lower greenhouse gas forcing, (iii) reduced ocean volume (freshwater transferred to the expanded ice sheets), and less carbon in the terrestrial biosphere (implemented by instead adding carbon to the atmosphere + ocean). The duration of this new, glacial spin-up needs to be sufficient to bring the system into (a new) equilibrium (of which the slowest adjusting component will be sediment composition (wt% CaCO$_3$), although continuing changes in wt% CaCO$_3$ will to some extent be reflected in continuing changes in pCO$_2$ (why?)).

Before carrying on, check that everything is 'correct' (or at least: understandable) so-far:

* Confirm that ocean salinity is now higher than the preindustrial *restart* that was originally provided. The file `biogem_series_ocn_sal.res` is the time-series results file for ocean salinity – the 2nd column is the mean ocean salinity. Originally it was 34.904 PSU (or ‰), now it should be about 35.9.
  Note how atmospheric pCO$_2$ has responded to the change in ocean volume and sea-level (and tracer concentrations) alone (you did try testing the 4 different glacial boundary conditions separately first, right, and hence have 4 experiments in which you have fully quantified their individual effects ... ?). How does this reduced resolution version of *c*GENIE compare to published estimates (too much; too little; why? or if 'about right' – does this mean that you can completely trust *c*GENIE from now on?).
* Confirm that you have a colder ocean (due to altered albedo **and/or** greenhouse radiation forcing) ... no, seriously! You never know what might have gone wrong with a simple slip of the keyboard ... Surface ocean temperature also has established proxies for its glacial value and so the model can be contrasted against data.
  The file `biogem_series_ocn_temp.res` is the time-series results file for ocean temperature – the 2nd column is the mean ocean temperature, the 3rd column is mean sea surface temperature (SST), and the 4th mean benthic (deep (> 2 km) ocean floor). How much colder has it become? Is this realistic? Analyze the SST distribution (the surface field of the 3D netCDF time-slice file, or the `'sur_*'` variables in the 2D netCDF file) – how does this compare to observations? In the book chapter the data-based difference in SST between the LGM and Holocene is given. However, the map given is for the glacial-interglacial difference, which happily is something you have previously learned to do using Panoply (I hope!).
  Also, what is the CO$_2$ impact of lower SSTs? Note that you may not be able to directly compare your CO$_2$ prediction with all previous studies (e.g. summarized in the book chapter) because in your model, sea-ice and ocean circulation will also have been affected to some extent by the climate change. (How much have they been affected? There are also some proxies for sea-ice extent as well as ideas and hypotheses about ocean circulation changes.) Many (but not all) previous model studies have simply estimated CO$_2$ changes due to temperature change with fixed sea-ice and circulation fixed, by prescribing a different ocean surface temperature for the CO$_2$ solubility calculation. (This is a little beyond the scope of what you are expected to do here, but can be done in GENIE.)
* Confirm that you have correctly added (rather than subtracted!) carbon to the ocean+ atmosphere. The ocean + atmosphere carbon inventories should start changing from the start of the experiment incorporating the carbon change forcing (`p0000b_FeMahowald2006_ADJUST_terrestrialC`) and the change should be approximately uniform. You can calculate the change in ocean + atmosphere carbon inventory from the atmospheric CO$_2$ time-series file (`biogem_series_atm_pCO2.res` – column #2 is the global CO$_2$ inventory in mol) and the ocean total dissolved carbon time-series file (`biogem_series_ocn_DIC.res` – column #2 is the global DIC (total dissolved inorganic carbon) inventory in mol). Note you will have to convert from mol to gC (or PgC) in order to compare to the amount you requested. If the rate of inventory change turns out to be not quite linear, and particularly if the inventory change should turn to be not quite what you were expecting ... why? (Hint: refer to the mechanisms discussed in the lecture

(and papers) relating deep-sea sediments and weathering to changes in total carbon (e.g., fossil fuel CO_2 release.)

It will be important to your subsequent understanding and also investigations into low glacial CO_2, to spend some time ensuring you understand what is going on when carbon is added or subtracted to the ocean in a model that accounts for the additional longer time-scale $CaCO_3$ sedimentary processes and weathering. It might help to also follow the time-series file of mean surface sediment composition (wt% $CaCO_3$) – `biogem_series_sed_CaCO3.res` and compare it to the evolution of atmospheric CO_2 and also total ocean+atmosphere carbon inventory. (You might also follow: `biogem_series_focnsed_CaCO3.res`, which is the flux of $CaCO_3$ to the sediments, and `biogem_series_fsedocn_Ca.res`, which is the sediment $CaCO_3$ dissolution flux. The difference between them is obviously the global burial rate of $CaCO_3$ – how does this compare with the prescribed weathering rate (parameter: `rg_par_weather_CaCO3`), which is set to a value of: `0.859104E+13` mol yr^{-1} in the *user config* files provided.)

A2.2 Unless you are extremely lucky and already have a value of atmospheric CO_2 that is 90 ppm lower than the modern spin-up (pre-industrial == 278 ppm) ... (wtf?!) you may want to test other changes that might have taken place between glacials and interglacials that affected CO_2. Obviously a spot of creating of new *user-config* files will be in order here (perhaps using: `exp12_glacial` as a template, but it is entirely up to you). Ideally, you would test the impact of each change underlined individually first before combining them, so as to develop a better understanding of the different ways in which CO_2 is controlled (and the associated impacts on other elements of the global carbon cycle and climate) underlined before bunging everything in together.

Some underlined suggestions (i.e., not an exhaustive list, nor a prescribed one (i.e., not everything necessarily has to be done!) list, follow:

- **Global weathering rate**. Refer to *Ridgwell and Zeebe* [2005] for the role of weathering. Also to *Kohfeld and Ridgwell* [2009] for some references to the changes in weathering that might have taken place between glacial and interglacial. The namelist parameter that controls the annual rate of solute input into the ocean is:
  `rg_par_weather_CaCO3=0.9E+13`
  Either edit this value (under heading: # --- WEATHERING ---) or add a new line at the end of the *user config* file specifying the value you want. Units are mol of $CaCO_3$ weathered per year.
  This parameter could also be adjusted to implicitly simulate the effect of a change in carbonate deposition in coral reefs and other shallow water carbonates, changes that GENIE cannot simulate explicitly. See *Ridgwell et al.* [2003] (http://www.seao2.org/pubs/ridgwell_et_al_2003a.pdf) for references and discussion of the sort of change in carbonate deposition on the shelves that might have taken place. A decrease in $CaCO_3$ removal on the continental shelves can be simulated by increasing the weathering flux to the open ocean. In other words, you can look at the parameter `rg_par_weather_CaCO3` as representing the residual weathering flux to the open ocean, after some of the weathering flux has been removed in coastal areas. Even if global weathering of the continents did not change, any reduction in $CaCO_3$ precipitation and removal on the continental shelves would result in an increased solute flux to the open ocean.
- **Iron fertilization**. Read up on this first, e.g., see references in *Kohfeld and Ridgwell* [2009]. The glacial was dustier than present, hence there can only have been increased aeolian iron supply to the ocean surface. However, what is not so clear is how important (relative to Fe being upwelled) aeolian Fe is today, let alone during the last glacial ...
  Anyway: one way to increase the aeolian Fe supply to the ocean surface is simply to increase the solubility of the Fe in dust. This is controlled by the parameter:
  `bg_par_det_Fe_sol=0.0015`
  with the default being a global average dust Fe solubility of 0.15% (fraction == 0.0015). Increasing will increase the Fe input to the ocean surface everywhere (in direct proportion to the modern spatial pattern). The pattern of total aeolian Fe supply is recorded in the (2D BIOGEM) variable: `misc_sur_fFetot_mol`, with the dissolved component under:

`misc_sur_fFe_mol` (`misc_sur_Fe_sol` is the map of solubility, which in GENIE is not uniform in space – any idea what the reason for this assumption might be?).
(A glacially-explicit map of dust deposition could also be applied in place of the modern deposition map – if you would like to test this, I can create one, but note that there are very significant 'errors' in re-gridding dust maps to this highly simplified continental topography.)

- **Remineralization depth**. There is no temperature control on the rate of bacterial degradation of sinking organic matter (see: book chapter + references therein) but the effect of lower ocean temperatures and a slower rate of bacterial degradation of organic matter can be simulated by specifying that particulate organic matter reaches greater depth before being remineralized (and CO_2 and PO_4 released back to the seawater). The namelist parameter that controls the e-folding depth reached by particulate organic matter before remineralization is:
`bg_par_bio_remin_POC_eL1=589.9451`
Either edit this value (under heading: # --- REMINERALIZATION ---) or add a new line at the end of the *user config* file specifying the value you want. Units are m.
Read *Ridgwell et al.* [2007] for additional discussion of this parameter. See Figure 2-4 in *Ridgwell* [2001] (http://www.seao2.org/pubs/ridgwell_thesis.pdf) for an illustration of how the flux of particulate organic matter decreases with depth in the ocean, plus references therein.
There is also an associated parameter: `bg_par_bio_remin_POC_frac2`, which sets a fraction of organic matter that is assumed to settling through the water column completely un-altered (currently assigned a value of `0.025` == 2.5%), but this is arguably less appropriate to change than the remineralization length-scale of the more labile fraction (97.5% of exported particulate organic carbon).

- **Macro nutrient inventory and uptake**. Suggestions have been made that nutrients were used more efficiently during the LGM, meaning that for the same nutrient uptake at the surface more carbon was exported to depth in the ocean. See: *Omta et al.* [2006]. There are also a bunch of (relatively old) hypotheses concerning differences between glacial and modern ocean in how much nitrate (NO_3^-) there was. There is no NO_3^- in this version of GENIE (just PO_4^{3-} and Fe), but an analogous change can be made to the phosphorous cycle.
For the nutrient-to-carbon ratio in organic matter, the relevant parameter is:
`bg_par_bio_red_POP_POC=106.0`
To change the default value (106.0), add a new line at the end of the *user-config* file specifying the value you want. A larger number means that PO_4 is being utilized more efficiently and more organic matter ir being produced for the same nutrient consumption.
If you would like to test the effect of adding more PO_4 to the (glacial) ocean – a forcing is provided, called:
`p0000b_FeMahowald2006_ADJUST_phosphate`

- **CaCO3:POC rain ratio**. Kicked off by a classic 1994 Nature paper by Archer and Maier-Reimer (see: *Kohfeld and Ridgwell* [2009]), one powerful means of changing atmospheric CO_2 that has been proposed involves changes in the export ratio between $CaCO_3$ (shells) and POC (particulate organic matter). Such a change in ratio could come about through a variety of ways (e.g., via the 'silica leakage hypothesis' (see: *Kohfeld and Ridgwell* [2009]) and also through the direct effect of Fe on diatom physiology (see *Watson et al.* [2000] in Nature and also Supplemental Information). There are also ideas about an opposite ocean acidification effect, whereby the <u>less</u> acidic glacial (compared to modern) ocean led to <u>increased</u> calcification and $CaCO_3$ export.
In GENIE, the $CaCO_3$:POC rain ratio is controlled (technically: scaled) by the parameter:
`bg_par_bio_red_POC_CaCO3=0.03`
The pattern of $CaCO_3$:POC rain ratio is not uniform across the ocean (why? (see: *Ridgwell et al.* [2007, 2009]), and its pattern can be viewed in the (2D BIOGEM) netCDF variable:
`misc_sur_rCaCO3toPOC`.

- **Sea-ice extent**. Changes to sea-ice extent have already taken place due to changes in radiative forcing and planetary albedo (made previously). There is no much you can do to

further adjust sea-ice extent, other than via further changes to climate (via radiative forcing and/or albedo).

- **Atlantic circulation**. There are a variety of ideas and hypotheses about glacial ocean circulation and what influence it had on atmospheric CO_2. At least with respect to making tests and experiments in models, a common ploy has been to produce a collapsed AMOC (e.g., see *Chikamoto et al.* [2008] (*JGR* **113**)). Rather than apply a continuous freshwater forcing to the ocean throughout an extended (sediment interaction) time-scale (why would this not be a good idea?), there is a parameter in the model which creates an adjustment of the salt balance between the different ocean basins (to make the Atlantic more salty compared to the Pacific). (In other words: salt/freshwater is re-partitioned between the ocean basins rather than 'new' freshwater or salt externally added.) This parameter is:
  `ea_28=0.726862013339996340`
  Setting it to e.g., 0.0, will result in a collapsed AMOC. But maybe that is too extreme? (You might read up a little on the glacial ocean circulation literature and chose a value that gives as an appropriate change to Atlantic circulation as you can judge from the data and literature.)
- **Global ocean circulation / 'brine rejection'**. Some recent research has focussed on the possible role of 'brine rejection' in creating a saltier Antarctic bottom waters (e.g. see Adkins et al. 2002 Science paper) and hence a denser and more stratified deep ocean,, with the idea being this will trap carbon more efficiently. For a very recent study (and references therein), see:
  http://www.clim-past.net/6/575/2010/cp-6-575-2010.html
  GENIE has the capability to include this effect (at least crudely) and similarly to *Bouttes et al.* [2010]. For this, three namelist parameter values need to be set:
  `bg_ctrl_force_GOLDSTEInTS=.TRUE.`
  `bg_par_misc_brinerejection_frac=0.1`
  `bg_par_misc_brinerejection_jmax=9`
  The first, simply allows the BIOGEM biogeochem module to directly influence ocean circulation. The second is the fraction of salt, rejected during sea-ice formation (e.g., see *Bouttes et al.* [2010]) that is transferred directly to the bottom-most (underlying) ocean cell in the model. The first sets a latitude limit (counted in cells) to the effect – a value of 9 will restrict brine rejection to the Southern Ocean; a value of 18 will allow it to take place in the North Atlantic as well. (Note that in e.g., *Bouttes et al.* [2010], the effect is considered only in the Southern Ocean.)
- **MISC**. There are of course other possibilities for adjusting the model, although you need an a priori reason for doing so and what about the possible glacial state of global carbon cycling and climate you are trying to encapsulate. Examples might include wind speed (or air-sea gas exchange). Also note that uncertainty associated with changes in the terrestrial biosphere – if you can justify it, you are at liberty of course to chose values rather different to the default (500 PgC – implemented in the forcing as 1 PgC per year over 500 years).

A2.3 Commonly in (glacial CO_2) modelling studies, a steady state (or quasi steady state) simulation is run for the glacial (and compared to pre-indsutrial). The version of GENIE you have is sufficiently fast to do this quite effectively. It is possible to do non-state (glacial-interglacial) simulations, e.g. *Ridgwell* [2001], but this is rather more involved.
Note that in all of the above possible adjustments to the global carbon cycle, the mechanism of carbonate compensation is operating. Hence there will be direct (changes in carbon cycling within the water column) and indirect (interaction between ocean and deep-sea sediments) processes operating that will affect CO_2. Carbonate compensation will typically take a few 10s of thousands of years to fully adjust atmospheric CO_2. Not all previous modeling studies include this effect and in some cases it can drastically influence the predicted change in atmospheric CO_2.

A2.4 Even if you achieve atmospheric CO_2 of ca. 190 ppm (and actually, with some mechanisms on their own and also in combination, it is quite easy to achieve this), how do you know if you are 'right'? Many of the important constraints are summarized in *Kohfeld and Ridgwell* [2009] and *Archer et al.* [2000]. In particular:

- The distribution of the $CaCO_3$ content of deep-sea sediments. e.g., see Figure 6 in Archer et al. [2000]. You are not 'allowed' to blanket the entire ocean floor with $CaCO_3$ if you want to be consistent with the paleoceanographic record of the LGM ;)
  The predicted distribution of the $CaCO_3$ can be used to assess your circulation change – note that there is much less $CaCO_3$ in sediments in the North Atlantic at the glacial [*Archer et al.*, 2000]. See: *Chikamoto et al.* [2008] for a model assessment of the impact of AMOC changes on deep-sea sediment composition.
- The ocean should not go 'anoxic' (i.e., little to no dissolved oxygen left) over large expanses. (But you might consider this relative to the modern configuration – i.e., should the modern simulation under-estimate oxygen concentrations in the deep ocean, so will the glacial simulation, even if you get the mechanisms exactly 'right'.)
- There is a map of estimated changes in the biological flux to the ocean floor in *Kohfeld and Ridgwell* [2009] (also read the original reference). In the 2D netCDF file, the variable `focnsed_POC` gives you the flux of particulate organic matter (actually, carbon) to the ocean floor. By constructing a difference map of your glacial-interglacial predicted changes, you could contrast directly to the *Kohfeld et al.* [2005] reconstruction.
- The GENIE model is set up to predict $\delta^{13}C$ distributions. See: *Curry and Oppo* [2005]. There is also an atmospheric record of $\delta^{13}C$ (also predicted by GENIE) – see: *Smith et al.* [1999] and a more recent paper in GBC: *Lourantou et al.* [2010] ('Constraint of the CO_2 rise by new atmospheric carbon isotopic measurements during the last deglaciation ').
- Other proxies offer varying constraints at the global or regional scales. e.g., see: *Elderfield and Rickaby* [2000] (Cd/Ca ratios).

A2.5 REMEMBER: It is up to you whether you aim for ~190 ppm, or want to retain as much consistency with other (paleoceanographic) constraints as possible. Contrast your glacial CO_2 problem solution both with observational constrains as well as previous model studies. Discuss the appropriate caveats to your work (model deficiencies, uncertainties in data constraints). Make recommendations to enable future advances in understanding the glacial carbon cycle(?)