A Hitchhikers Guide to the *advanced* Black Arts
(of Earth system modelling)

VII: '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

13. Accounting for low glacial CO$_2$

13.0   Your task is to explain low glacial atmospheric CO$_2$.

It is up to you whether you aim for ~190 ppm, or want to retain as much consistency with other (paleoceanographic) constraints as possible.

In addition to the control and spin-up *user-configs*, you have been provided with: `exp12_glacial` which can be used as a template *user-config* file for you to base (if you want!) your glacial CO$_2$ investigations on (at least initially).

13.1   Before embarking upon some glacial experiments (below) – other model outputs/fields you might also keep in mind and for which some data/observational constraints on your glacial CO$_2$ 'solution' may be available (e.g. see *Kohfeld and Ridgwell* [2009]), include:

- The 2D `fields_biogem_2d.nc` field: `phys_seaice` ('sea-ice cover'), for which some glacial sea-ice limit information exists. (CLIMAP)
- 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. (CLIMAP, MARGO)
- 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. (Old Catubig paper in GBC)
- 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 δ^{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. (Jess Atkins Science paper)
- 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. (Cd/Ca proxies – e.g. papers by Elderfield and Rickaby)

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 – the above suggestions/guidance are intended as a starting point only.

13.2   In your glacial CO$_2$ investigations, 2 separate initial modifications of the model will nudge it in the vague direction of a glacial state (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 *c*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$_2$O 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 glacial boundary conditions are implemented as follows:

- To achieve a pseudo-glacial planetary albedo modification, add the following lines to a *user-config* file (and see **Lab V**):
  ```
  # 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, add the lines:
  ```
  # 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
  ```

If you like – you can 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. Ideally, these lines should ultimately be included in all (glacial) experiments that you require glacial albedo and radiative forcing for.

By all means play around with the albedo and radiative forcing climate boundary conditions, although one might wonder the reasoning behind adjusting radiative forcing below e.g. that appropriate for full glacial conditions. However, the planetary albedo is less certain as implemented in the model. For instance, one might legitimately adjust this to achieve appropriate last glacial sea surface temperatures (SST), or rather: a glacial-interglacial difference in SSTs similar between model and data.

13.3 By this point, you should have created a new (mostly) spun-up model state, incorporating: (i) higher planetary albedo, and (ii) lower greenhouse gas forcing. 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 $p CO_2$ (why?)).

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

13.4 Unless you are *extremely* lucky and already have a value of atmospheric CO_2 that is 90 ppm lower than 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 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 <u>individually</u> 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) <u>before</u> bunging everything in together.

Before diving straight in – note the number of different modifications of the global carbon cycle that might be considered and tested in the model. What is your methodological strategy going to be? Are you going to add all of the modifications into a single run and hope that you can understand what has happened at the end? (I hope not!) Are you going to run all the modifications individually (how?). Are you going to try combinations to test whether any combine non-linearly? You will want to make use of the cluster queue and submit at least some of the experiments. You will also need t already have a good idea of how long to run them before (hopefully you obtained this knowledge from the idealized perturbation experiments in **Lab VI**).

Some <u>suggestions</u> (i.e., this not an exhaustive list, nor a prescribed one and not everything necessarily has to be done!):

- **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.
  [**HINT**: keeping track of how mean sediment wt% $CaCO_3$ changes will be helpful, as may 2D sediment distributions and ultimately, the sediment core records.]

- **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.)
[**HINT**: viewing maps of particulate organic carbon (POC) export may be particularly helpful.]

- **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).
  [**HINT**: viewing distributions of PO_4 and/or O_2 in the ocean may be helpful. Perhaps also $\delta^{13}C$.]

- **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`
  Note that adjusting the ocean PO_4 inventory should only be done one (and not accidentally in each successive experiment!).
  [**HINT**: viewing distributions of PO_4 and/or O_2 in the ocean may be helpful. Also ocean sediment $CaCO_3$ distributions.]

- **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. Note that this response (higher saturation == great calcification) is encoded into your model configuration – see *Ridgwell et al.* [2007b]. 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`.
[**HINT**: viewing sediment $CaCO_3$ distribution may be helpful.]

- **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.)
[**HINT**: viewing distributions of PO_4 and/or O_2 and/or $\delta^{13}C$ in the ocean may be helpful. Also ocean sediment $CaCO_3$ distributions.]

- **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.)
[**HINT**: viewing distributions of PO_4 and/or O_2 and/or $\delta^{13}C$ in the ocean may be helpful. Also ocean sediment $CaCO_3$ distributions.]

- **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).

13.5 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).

Note that commonly in (glacial CO_2) modeling studies, a steady state (or quasi steady state) simulation is run for the glacial (and compared to pre-indsutrial). The version of *c*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.

Also 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.

13.6 Finally … maybe you have achieved close to 190 ppm and are not unreasonably consistent with various paleoceanographic proxies and are hence feeling rather pleased with yourself … O – sorry – I forget to mention a little something:

- There is as 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 associated with the last glacial.

Ahhhh … and I also forgot:

- It is thought that there was a 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 and hence predict this, but you are not using a version with this science module enabled.)

Unfortunately – both these effects act to increase atmospheric *p*CO_2 at the last glacial (or decrease it across the deglacial transition, which ever way around you like to think of it). So you are actually rather further off achieving a net 90 ppm difference than you thought.

To test the effect of a ~3% increase in salinity at the last glacial, you need to create a new *user-config* file (copied from your last glacial 'best guess' configuration), add the lines:
```
bg_ctrl_force_GOLDSTEInTS=.true.
bg_par_forcing_name='p0000b_FeMahowald2006_ADJUST_salinity'
```
and run using your last glacial best guess as the restart.

For the reduction in terrestrial carbon storage, you need to try a further experiment (on top of and/or following on from the salinity change experiment), with the *user-config* line:
`bg_par_forcing_name='p0000b_FeMahowald2006_ADJUST_terrestrialC'`

These two *forcings* 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 ...). Similarly with 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). 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!)

If you want to check that you have applied these *forcings* correctly:

- For the first – mean ocean salinity should end higher than the preindustrial *restart* that was originally provided (or compared to your best guess glacial run). 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. 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?).

- To confirm that you have correctly added (rather than subtracted!) carbon to the ocean+ atmosphere – thee 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.)

Glacial-interglacial CO₂ components