The Changing North Atlantic Carbon Sink

Galen A. McKinley[1]*, Val Bennington[1], Stephanie Dutkiewicz[2], David Ullman[1], and Dierk Polzin[1]

[1] Atmospheric and Oceanic Sciences, University of Wisconsin – Madison, Madison, WI
[2] Earth, Atmospheric and Planetary Sciences, Massachusetts Institute of Technology, Cambridge, MA
*galen@aos.wisc.edu

MOTIVATION

Previous work indicates:

- The North Atlantic is an important carbon sink, absorbing 23% of anthropogenic carbon stored in the oceans (Sabine et al., 2004), yet basin-integrated CO_2 flux variability is small (McKinley et al., 2004).
- Changes in uptake (Shuster and Watson, 2007) and interior carbon storage (Feely et al., 2005) may be significant.
- Variability in the spring bloom has a complex relationship to physical variability (Follows and Dutkiewicz, 2002) and biological impacts on carbon fluxes are poory understood.

We ask:

- What is the sign and magnitude of recent variability and trends in the North Atlantic carbon sink? What are the dominant controls?
- Does biological variability have a significant impact on the N. Atlantic carbon sink?

MODEL DESCRIPTION

PHYSICAL MODEL:

- MITgcm (Marshall et al. 1997a,b) configured for North Atlantic domain
- Horizontal resolution of 0.5° x 0.5°, 23 vertical levels ranging in resolution from 10m near surface to 500 m at depth
- KPP mixing layer scheme, GM-Redi representation of isopycnal mixing

BIOGEOCHEMICAL MODEL:

- Cycling of carbon, phosphorus, silica, iron, oxygen and alkalinity
- Ecosystem model includes 2 phytoplankton functional types, one zooplankton type and explicit DOM and POM (Dutkiewicz et al., 2005)
- Surface carbonate chemistry (Follows et al., 2006)
- Iron chemistry (Parekh et al., 2004)
- Air-sea exchange of carbon dioxide and oxygen (Wanninkhof, 1992)
- Atmospheric pCO_2 from Mauna Loa observations

MODEL VERIFICATION

- Run 1 - Spin up: 80 years physics only (daily NCEP)
    26 years physical-biogeochemical model (daily NCEP)
    Interannual: 1992 to 2006 (daily NCEP, relaxation to satellite-derived SST (Reynolds et al, 2002))
- Run 2 - Spin up: 81 years physics only (daily NCEP)
    20 years physical-biogeochemical model (daily NCEP)
    Interannual: 1981 to 2006 (daily NCEP)
- In both runs, the model simulates the circulation, sea surface temperature, salinity, and mixed layers reasonably well.
- The ecosystem model captures the timing and magnitudes of the North Atlantic bloom. The biomass in the subtropics is, however, too low, as is typical of this class of models.
- Model pCO_2 compares favorably to that observed by Takahashi et al. (2002). The model also captures the pCO_2 components due to temperature and non-temperature effects (Figure 1).
- Comparison to observations at Bermuda (Bates, 2007) indicate the model is capturing variability in surface pCO_2 quite well.
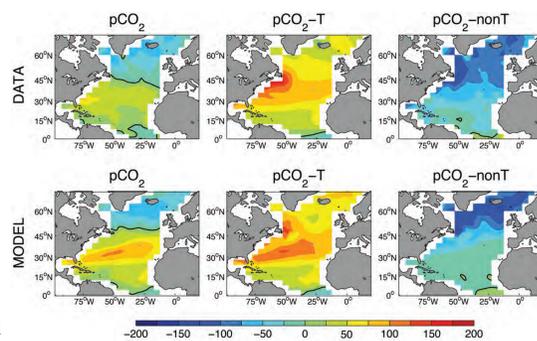
Figure 1: Amplitude of North Atlantic pCO_2 cycles (ASO-FMA). Takahashi et al. (2002) climatology (top) and MODEL (bottom). Temperature (pCO_2-T) and non-temperature (pCO_2-nonT) components are calculated following Takahashi et al., 2002. (Run 2)

CO_2 FLUX TREND, 1992-2006

- The basin-scale air-sea flux is increasing. The flux is also increasing in each gyre.
- Basin-wide, for 1992 and 2006, we find a sink increase of 0.09 PgC/yr/decade (Run 1). For the same period, Run 2 has a larger trend of 0.2 PgC/yr/decade.
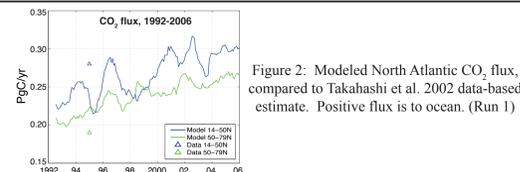- Results are consistent with the latest data-based estimate (Takahashi et al., 2002).

Figure 2: Modeled North Atlantic CO_2 flux, compared to Takahashi et al. 2002 data-based estimate. Positive flux is to ocean. (Run 1)

CONTROLS ON pCO_2 VARIABILTY

Subpolar Gyre: DIC MIXING

- The negative correlation between pCO_2 and SST indicates colder SST is associated with more mixing, and vice-versa.
- Mixing modifies DIC, driving pCO_2 change.
- ⇑SST → ⇓ MLD → ⇓ upwelling DIC → ⇓ pCO_2

Subtropical Gyre: Temperature

- The positive correlation between pCO_2 and SST indicates that SST anomalies directly influence pCO_2 via the thermodynamic effect on gases.
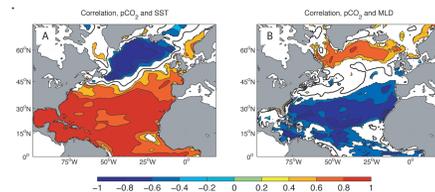- ⇑SST → ⇑ pCO_2 (thermodynamic effect on gases)

Figure 3: Correlation plots (a) pCO_2 and SST, (b) pCO_2 and MLD. Based on fields detrended and 12-month smoothed. Shaded regions are significant at the 95% confidence level. (Run 1)

DOMINANT PATTERNS OF pCO_2 TREND

- The first EOF of the pCO_2 (Figure 4a) explains 73% of the interannual variance, and its PC1 (Figure 4, inset) is characterized by a steadily increasing trend.
- EOF1, the dominant pattern of pCO_2 trend, illustrates greater increase of pCO_2 in the east and south (5-10 µatm), and slower increase in the west (0-5 µatm). These change are consistent in sign with recent changes observed in-situ (Schuster and Watson, 2007) though smaller in magnitude.
- Regression of pCO_2 components (pCO_2-ALK, pCO_2-DIC, pCO_2-T) onto PC1 unveils localized regions of high component variability
- Warming and freshening in the Labrador Sea waters increase pCO_2, while the associated decrease in DIC decreases pCO_2.
- There is evidence of modified advection of high-DIC and high-ALK eastern subtropical waters along the NAC into the western subpolar region, consistent with observed spin-down of the subpolar gyre (Häkkinen and Rhines, 2004).
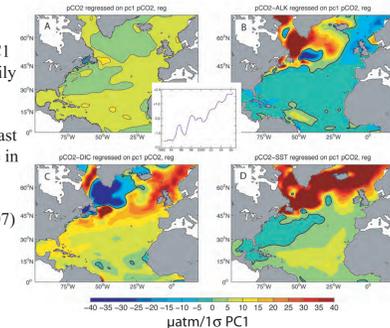
Figure 4: (a) EOF1 of pCO_2, and regression of (b) pCO_2-ALK, (c) pCO_2-DIC, and (d) pCO_2-T onto PC1 (inset). (Run 1)

BLOOM DOMINATES pCO_2 VARIABILITY

- Interannual variability of pCO_2 is highest during the spring/early summer in the subpolar gyre. This variability is driven by year-to-year changes in pCO_2-nonT (Figure 5) associated with timing of the spring bloom.
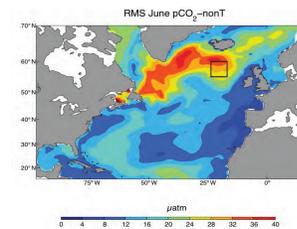- The seasonal cycle of pCO_2 at Iceland (boxed region in Figure 5) has strong, opposing impacts from temperature and DIC-driven influences (Figure 6).
- The dominance of biological variability to surface ocean pCO_2 variance suggests that bloom variability may be a first order control on annual CO_2 flux variability in the subpolar gyre.

Figure 5: Standard deviation of June pCO_2-nonT between 1982 and 2006. (Run 2)
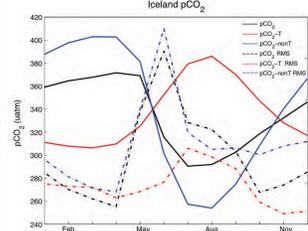
Figure 6: Climatology and RMS of pCO_2 and its temperature and non-temperature components within the 5°x5° box at Iceland (57.5°N,17.5°W). (Run 2)

... BUT DOES NOT TRANSLATE TO CO_2 FLUX

- DIC controls interannual variability of pCO_2 in the subpolar gyre and is significantly anti-correlated with annual CO_2 flux variability (Figure 7, see also Figure 3).
- However, change in annual carbon export is not associated with CO_2 flux changes in most of the subpolar gyre (Figure 8); i.e. biological variability is not driving CO_2 flux variability.
- In the small region centered at 60N, 30W, there is a correlation between CO_2 flux and export, but it is indirect. Here, enhanced winter mixing increases surface DIC and increases pCO_2, lowering the flux; and at the same time weakens the bloom.
- Correlations between export and CO_2 flux in the subtropical region are also indirect. Figure 3 illustrates that SST drives pCO_2 change in the subtropics. Colder surface temperatures are associated with more mixing, and thus greater nutrient supply, and vice-versa.
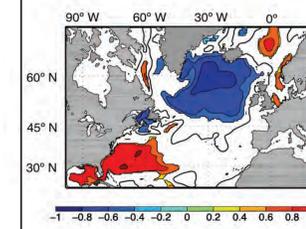
Figure 7: Correlation between anomalies of annual surface DIC and annual CO_2 fluxes. Shaded regions are significant at the 95% confidence level. (Run 2)
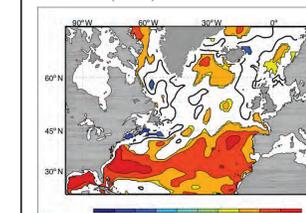
Figure 8: Correlation between anomalies of annual export of POC below 100m and CO_2 fluxes. (Run 2)

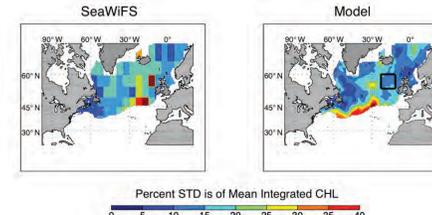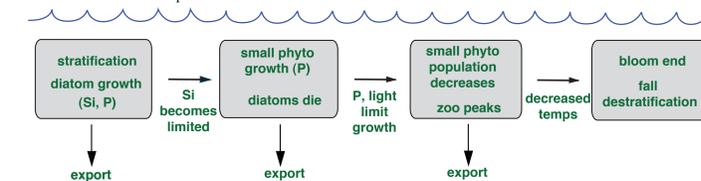## FLUX IMPACT SMALL BECAUSE BLOOM STRENGTH VARIES ONLY 10-25%

- Annually integrated chlorophyll in the subpolar North Atlantic varies by an average of 15% in the model and 14% in SeaWiFS satellite data between 1998 and 2006 (Figure 9).
- Blooms that start earlier tend to end earlier, as seen in the 25-year timeseries of POC export at Iceland (Figure 10).
- Modeled variability of annually-integrated export is smaller than chlorophyll variability (5-20 %).
- Thus, we find that variability in the biological pump of carbon is only 5-20% across the North Atlantic.

WHAT LIMITS VARIABILTY IN THE BIOLOGICAL PUMP?

- Blooms in the subpolar region begin with stratification. Diatoms bloom first, using silicate and phosphorous.
- When silicate becomes deplete, the ecosystem becomes dominated by small phytoplankton, using the remaining phosphorous.
- When nutrients and/or light becoming limiting, potential growth is overcome by mortality and grazing. The bloom ends.
- This succession was observed during NABE (Sieracki et al., 1993; Lochte et al., 1993), over one seasonal cycle. We find that the same sequence limits interannual variations.

Figure 9: Integrated chlorophyll variability 1998-2006 as a percent of the annual mean chlorophyll SeaWiFS (left) and MODEL (right). (Run 2)

Figure 10: 25 years of export of POC at 100m at Iceland for model (1982 - 2006). Climatology in black. (Run 2)

stratification → diatom growth (Si, P) → Si becomes limited → small phyto growth (P) diatoms die → P, light limit growth → small phyto population decreases zoo peaks → decreased temps → bloom end fall destratification

export → export → export

SUMMARY

- Interannual variability in pCO_2 and CO_2 flux is controlled by temperature in the subtropics and by DIC mixing in the subpolar gyre (Fig. 3, 7).
- For 1992-2006, pCO_2 across the gyre has increased 0-5 µatm in the west and 5-10 µatm in the east and south. Over the same period, changes in pCO_2-T, pCO_2-ALK and pCO_2-DIC can exceed ±40µatm north of 45° N. (Figure 4)
- Large summertime variability in subpolar gyre pCO_2 (Figure 5, 6) is driven by changes in bloom timing. But, because integrated bloom strength variability is small (Figure 9, 10), this does not translate to large changes in carbon export (Figure 8).
- We conclude that biological variability is not a first order control on variability in the North Atlantic carbon sink.
- Modeled patterns of pCO_2 trend (Figure 3a), driven physically, help explain trends observed in-situ (Shuster and Watson, 2007).
- Future work: Study of multi-decadal variability over the entire NCEP/NCAR Reanalysis period.

REFERENCES

Dutkiewicz, S., et al. 2005. Global Biogeochem. Cycles, 19, GB1021, doi:10.1029/2004GB002342.
Bates, N, 2007. J. Geophys. Res., 112, C09013, doi:10.1029/2006JC003759.
Bennington, V., G. A. McKinley, S. Dutkiewicz and D. Ullman, 2008. submitted to Global Biogeochem. Cycles.
Feely, R., L. Talley, G. Johnson, C. Sabine and R. Wanninkhof, 2005: EOS, 86, 42, 404-405.
Follows, M.J., and S. Dutkiewicz, 2002. Deep-Sea Res. II, 49, 321-344.

Feely, R.A,et al., 2005. EOS 86, 399, 404-405.
Häkkinen, S. and P.B. Rhines, 2004. Science, 304, 555-559.
Lochte, K., H.W. Duckow, M.J.R. Fasham and C. Stienen, 1993: Deep Sea Research II, 40, 91-114.
Marshall, J. C., C. Hill, L. Perelman, and A. Aderof, 1997a. J. Geophys. Res., 102, 5733 – 5752.
Marshall, J. C., A. Adcroft, C. Hill, L. Perelman, and C. Heisey 1997b. J. Geophys. Res., 102, 5753 – 5766.
McKinley, G.A., M.J. Follows, and J. Marshall, 2004. Global Biogeochem. Cycles, 18, GB2011, doi:10.1029/2003GB002179.

Parekh, P., M.J. Follows and E.A. Boyle, 2004. Global Biogeochem. Cycles, 18, GB1002, doi:10.1029/2003GB002061
Reynolds, R.W., N.A. Rayner, T.M. Smith, D.C. Stokes, and W. Wang, 2002. J. Climate, 15, 1609-1625.
Sabine, C., et al., 2004. Science, 305, 365-371.
Schuster, U., and A.J. Watson, 2007. J. Geophys. Res., 112 (C11), C11006, doi:10.1029/2006JC003941.
Sieracki, M.E., P.G. Verity and D.K. Stoecker, 1993: Deep Sea Research II, 40, 213-225.
Takahashi, T., et al., 2002. Deep-Sea Res. II, 49, 1601–1622.
Ullman, D., 2008: Masters Thesis, University of Wisconsin-Madison.
Wanninkhof , R., 1992. J. Geophys. Res., 97, 7373–7382.

April 2008