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Effect of lower sea level on geostrophic transport through the Florida Straits during the Last Glacial Maximum

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[1] We investigate the effect of a 120 m sea level drop on transport through the Caribbean Sea and the Florida Straits during the Last Glacial Maximum (LGM) relative to the present, using the Regional Ocean Modeling System (ROMS). A geostrophic transport estimate for the Florida Straits suggests the LGM Florida Current was weaker than today by one third assuming that the velocity at the bottom of the channel was as small as it is today. This is consistent with a decrease in the North Atlantic overturning circulation, but there are other possible reasons for a flow decrease. It is possible that a shallower LGM Florida Straits sill depth could cause the diversion of some flow from the Florida Current. Our model results show that the volume transport through the Florida Straits is slightly reduced in a lower sea level model simulation when compared to a control sea level simulation (34.8 ± 2.0 Sv versus 39.8 ± 2.3 Sv). The difference in transport is of the order of 5 Sv, likely representing a maximum limit to the LGM flow reduction due to sea level change. Therefore, the change in sill depth between the LGM and the present is unlikely to have been a cause of the entire observed flow reduction. We also use the model output to demonstrate that transport through the Florida Straits can be accurately calculated from ocean margin density data at the core locations in the work of Lund et al. (2006) and Lynch-Stieglitz et al. (2009), provided that a sufficiently deep reference level is chosen.

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1. Introduction

[2] The modern North Atlantic surface circulation (shallower than 1 km) consists of a wind-driven clockwise subtropical gyre superimposed on the surface component of the meridional overturning circulation (MOC), which flows northward along the western boundary and is compensated by a southward deep water flow.

[3] Water flowing westward and northward through the Caribbean Sea and the Florida Straits includes components of both the wind-driven gyre ($\sim$17 Sv) and the surface compensation for North Atlantic deep water export ($\sim$14 Sv). Both components enter the Caribbean Sea through the Antilles Islands channels, pass through the Yucatan Channel, and exit through the Gulf of Mexico and the Florida Straits [*Johns et al.*, 2002]. This adds up to $\sim$31 Sv net transport through the Florida Straits. Observations show 18.4 ± 4.7 Sv entering the Caribbean through the Lesser Antilles, 3.0 ± 1.2 Sv through the Mona Passage and 7–8 Sv through the Windward Passage [*Bulgakov et al.*, 2003], resulting in a total Caribbean inflow of 28.4 Sv [*Johns et al.*, 2002]. Approximately 3–4 Sv enter the Florida Straits from the NW Providence and Santaren channels [*Leaman et al.*, 1995], adding up to around 32 Sv of water flowing northward through the Florida Straits [*Baringer and Larsen*, 2001; *Hamilton et al.*, 2005], above a sill depth of about 760 m.

[4] During the last glacial maximum (LGM), approximately 21,000 years ago [e.g., *Bard et al.*, 1990; *Yokoyama et al.*, 2000], the North Atlantic MOC was markedly different from today. There is evidence of a shallower LGM overturning [e.g., *Curry and Oppo*, 2005], but the strength of the overturning has not yet been definitively established [*Lynch-Stieglitz et al.*, 2007].

[5] Since the δ^{18}O of foraminifera shells is a proxy for the density of the water in which they formed, geostrophic transport can be estimated from foraminifera δ^{18}O measurements on the sides of the Florida Straits. A geostrophic transport estimate using a reference level at the bottom of the Florida Straits suggested that during the LGM flow was weaker by one-third relative to today [*Lynch-Stieglitz et al.*, 1999b].

[6] While the authors noted that a reduced density gradient across the LGM Florida Current is consistent with a decrease in the Atlantic overturning circulation, there are other possible explanations for the data. The flow may not have decreased by this much if the current was more barotropic, i.e., there was significant flow near the bottom of the channel. Even if the flow did decrease, it could represent a change in the wind driven flow through the straits or a diversion of some of the western boundary flow outside of the straits. A lower LGM sea level might induce such a diversion, especially for the deeper components of the flow.

[7] Our focus here is limited to the effect of a shallower Florida Straits. It is estimated that sea level was lower by

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about 120 m during the LGM [e.g., *Alley et al.*, 2005; *Bassett et al.*, 2005; *Bard et al.*, 1990; *Lambeck and Chappell*, 2001], which reduced the Florida Straits sill depth of 760 m [*Malloy and Hurley*, 1970] to 640 m. The sea level drop could either have a negligible effect on the Florida Current transport – the flow increases either its vertical shear or bottom velocity, transporting the same amount of water as today – or block transport of water flowing below 640 m at present. Such a diversion would intensify the Antilles current. Since the modern-day Florida Current extends down to the sill, it is important to test whether the some of the reduction in the LGM Florida Current inferred by *Lynch-Stieglitz et al.* [1999b] may have been due to a shallower sill depth.

[8] In this study, we test the sensitivity of the volume transport through the Florida Straits to a sea level drop of 120 m. For this purpose, we use two runs of the Regional Ocean Modeling System (ROMS), one with modern-day bathymetry and another with the bathymetry raised by 120 m to simulate lower sea level conditions during the LGM. We also use the modeled data to test whether transport can be accurately reconstructed using ocean margin density as was done by *Lynch-Stieglitz et al.* [1999b, 2009] and *Lund et al.* [2006].

2. Methods

2.1. Model

[9] The Regional Ocean Modeling System (ROMS) is a high resolution, bathymetry-following ocean model. ROMS solves free-surface, hydrostatic, eddy-resolving primitive equations on a grid of stretched, terrain-following coordinates in the vertical and orthogonal curvilinear coordinates in the horizontal [*Shchepetkin and McWilliams*, 2005; *Haidvogel et al.*, 2008]. In the Intra Americas Seas (IAS) configuration the model grid has a horizontal resolution of 10 km with 30 vertical levels, more closely distributed toward the surface. The grid covers the Caribbean Sea, the Gulf of Mexico, and the Florida Straits. The model bathymetry is derived from the [*Smith and Sandwell*, 1997] bathymetric map with a 2 arc minute cell size. Smoothing is applied to minimize pressure gradient model errors associated with strong topographic slopes, however particular care is given to insure the sill depths of all important passages and straits are accurately represented.

2.2. Boundary Conditions and Forcing Functions

[10] The model has two open boundaries. At these boundaries, a radiation condition is applied to the model state variables (temperature, salinity, velocity and free surface) [*Marchesiello et al.*, 2003], along with nudging to monthly climatological conditions derived from a ROMS simulation of the North Atlantic (J. Levin et al., personal communication, 2005). Configuration parameters for the open boundary conditions and mixing are similar to the ones used in [*Di Lorenzo*, 2003] and [*Marchesiello et al.*, 2003]. At the surface the model is forced with climatological monthly mean wind stresses derived from the 2000–2004 QuickSCAT winds blended with the NCEP reanalysis [*Milliff and Morzel*, 2001].

2.3. Experiments

[11] We perform two model integrations of 16 years length each. The first integration uses the control (modern) bathymetry. For the second integration, the bathymetry is raised by 120 m to simulate lower sea level conditions of the LGM [*Peltier and Fairbanks*, 2006], while all other parameters remain unchanged. This approximate LGM bathymetry is likely to be relatively close to the actual bathymetry because the isostatic adjustment of the crust due to the transfer of water from the ocean to the large ice sheet on North America is minimal in the Caribbean region [*Peltier and Fairbanks*, 2006]. While uplift rates can be high in some of the tectonically active regions of the Caribbean (20 m kyr^{-1} at Barbados) [*Peltier and Fairbanks*, 2006], the parts of the Caribbean that are of greatest interest to us are not tectonically active and there is only modest subsidence ($2–3 \times 10^{-2}$ m kyr^{-1} at the Bahamas; 1.5×10^{-2} m kyr^{-1} in South Florida) [*Richards et al.*, 1994; *Toscano and Lundberg*, 1999]. In both runs, the bathymetry at the open boundaries is blended with the North Atlantic model bathymetry through a gradual transition over a buffer zone to ensure the same boundary mass flux.

[12] The mean sea level (MSL) of the model is MSL = h + *ssh_bar*, where h is the bottom topography and *ssh_bar* is the model time average sea surface height. The term *ssh_bar* is the dynamic component of sea level and is defined as fluctuations from a reference level of 0 m (Figure 1, MODERN). While *ssh_bar* is affected by the motion of the ocean, the reference depth of the water column depends only on $h(x, y)$ and is set by the gravitational field (e.g., the geoid) and the total amount of water volume. In the case of our LGM experiments we change the fixed reference by reducing the depth of the reference level by 120 m (Figure 1, LGM). To the extent that *ssh_bar* $\ll h$ we can treat the mean MSL as independent of water flow in our sensitivity experiments.

[13] We did not conduct experiments to test directly the sensitivity of model transports to different strengths of the overturning circulation. These experiments require different specifications of the mass flux at the model open boundary that cannot be realistically derived without data from a large-scale model of the entire North Atlantic that exhibit a slower overturning. Therefore our model experiments are limited to studying the sensitivity of flow in this region to changes in sea level alone.

3. Results and Discussion

3.1. Control Run

[14] The model stream function shows a net flow from the equatorial Atlantic Ocean into the Caribbean Sea and the Gulf of Mexico, and toward the North Atlantic through the Florida Straits (Figure 2), thus agreeing overall with observations [*Johns et al.*, 2002]. The control run indicates a total transport of 39.5 Sv through the Florida Straits, which is consistent both at the entrance from the Gulf of Mexico and at the exit into the North Atlantic. This transport is higher than the measured 32 Sv transport [e.g., *Baringer and Larsen*, 2001]. The model exhibits flow of 2.2 Sv out of

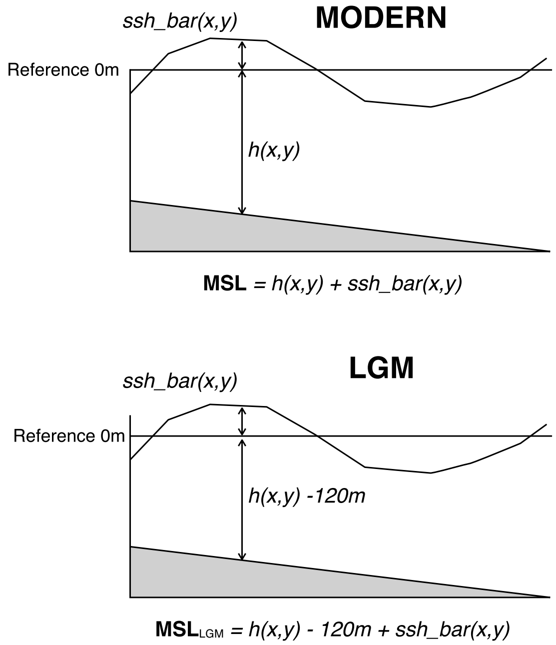

Figure 1. Schematic showing the treatment of the ocean mean sea level (MSL) in the MODERN and LGM experiments. In the top and bottom $h(x, y)$ is the depth of the water column with respect to a reference level of 0m (e.g., the geoid), and *ssh_bar*(x, y) represents the mean sea surface height of the ocean model determined by the mean ocean circulation.

the straits through the Old Bahama channel, compensated by an inflow of 2.5 Sv through the NW Providence channel, which is comparative to the 1.2 Sv observed inflow [*Leaman et al.*, 1995] (Figure 3 and Table 1). The Old Bahama channel outflow is opposite to the observed 2 Sv Santaren Channel flow into the straits [*Leaman et al.*, 1995].

[15] The volume entering the Florida Straits (39.5 Sv) is nearly balanced by the volume exiting the Florida Straits (39.8 Sv), and the 0.3 Sv difference is the difference between the Old Bahama Channel outflow (2.2 Sv) and the NW Providence Channel inflow (2.5 Sv). Transport measurements across numerous model sections indicate that transport tends to be conserved within 1%, an error that could be due to changes in the free surface elevation on time scales of a few days [*Hamilton et al.*, 2005].

[16] We find that the model transports in the control run are on average higher than the ones reported by observations. One possible explanation is that the model and observed transports are computed using averages over different period. Another possibility is that there are uncertainties in the model boundary conditions (e.g., atmospheric forcing, MOC boundary conditions, topography resolution, etc.) that lead to higher transports. In principal these boundary conditions can be corrected using data assimilation methods as shown in previous studies that use the ROMS modeling framework in the IAS to adjust the model boundary conditions to improve the model comparison with observations [*Powell et al.*, 2008]. However, to the extent that transports dynamics in the straits are not a strongly nonlinear function of the model boundary conditions (e.g., atmospheric forcing, MOC boundary conditions, topography resolution) we can still explore the linear sensitivity of the transports by varying only the sea level.

3.2. Sensitivity to Lower Sea Level

[17] The lower sea level run stream function has a similar appearance to the control (Figure 2), indicating that a 120 m sea level drop does not induce any major circulation change in the Caribbean Sea and the Florida Straits. There is a 37.1 Sv flow entering the Florida Straits between Cuba and the Dry Tortugas and 34.8 Sv exiting the straits into the Atlantic Ocean north of the Bahamas (Figure 3 and Table 1). The flow is around 2.5 Sv less than the control at the straits entrance and around 5 Sv less at the exit, with the difference exiting through the Old Bahamas (2 Sv) and the NW Providence (0.5 Sv) channels. This reduces the NW Providence channel inflow to 2.0 Sv and increases the Old Bahama channel outflow to 4.2 Sv. Given that the flow across the open boundaries does not change between the experiment and control run, the decrease in Florida Straits transport implies an increase in the Antilles current. These numbers suggest that the sill depth can have an effect on transport, albeit a small one for a sea level change of 120 m. Since the great majority of the transport is in the upper 600 m of the straits in our model (36.7 Sv according to Figure 4) just as in observations [*Leaman et al.*, 1995], it is not surprising that blocking the straits below 640 m would not impact the flow significantly. Since the model overestimates the modern flow through the Florida Straits, we view the 5 Sv reduction due to the bathymetric changes to be an upper limit.

3.3. Geostrophic Transport

[18] Geostrophic transport is first calculated across various sections throughout the straits based on the model's absolute geostrophic currents. The model geostrophic transport is calculated from the density and sea surface height of the model output and uses the same model numerics computing the pressure gradient term on the terrain-following coordinate system. The model geostrophic transport is very close to the model absolute transport (Table 1), supporting the notion that the transport through the Florida Straits is largely geostrophic.

[19] We then calculate geostrophic transport using two density profiles along the bottom layer (straits margins) on either side of each section as in [*Lynch-Stieglitz et al.*, 1999a, 1999b]. When the margin geostrophic transport is referenced to a zero velocity at the deepest part of the section, the geostrophic transport is underestimated in all but the deepest section (Section 1, Dry Tortugas). Using the bottom model layer velocity as a reference velocity brings the transports through all of the sections close to the actual transports. We can then conclude that using ocean margin density to estimate vertical density profiles yields an accurate calculation of the vertical shear in the geostrophic transport. However, an accurate total transport depends on the quality of the assumption about the reference velocity.

[20] In this model the bottom layer can be quite thick (75 m at Sections 3 and 5), and represents the average

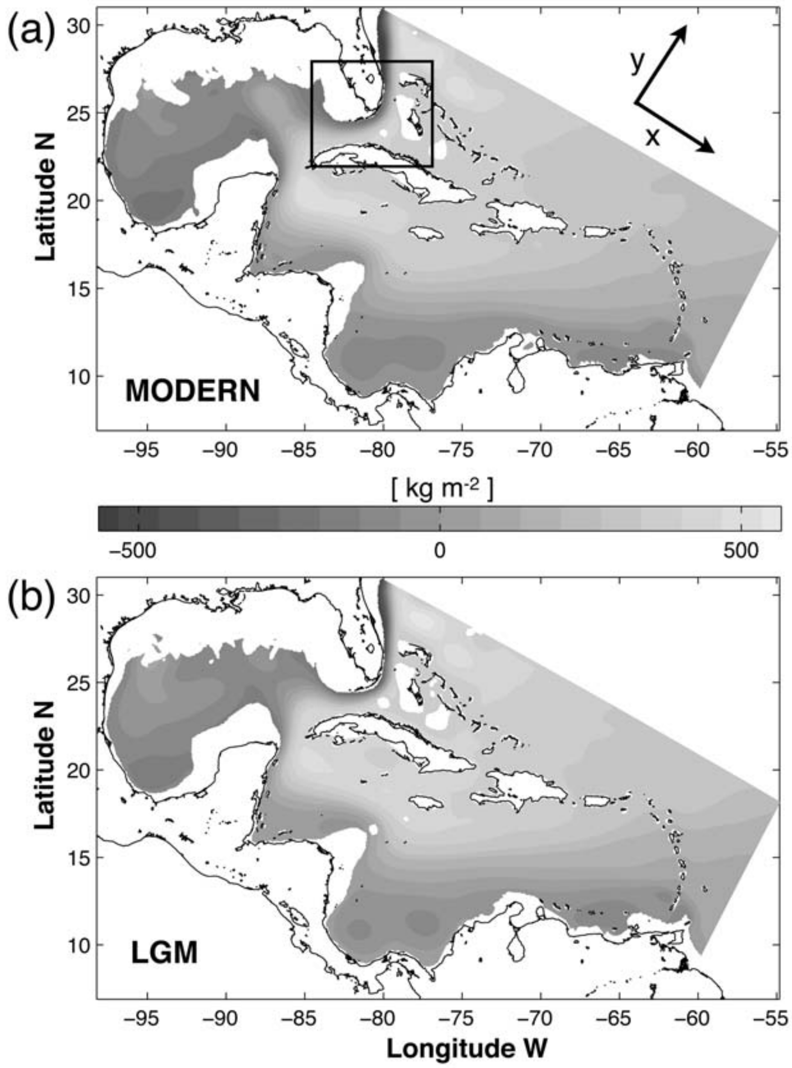

Figure 2. Stream function of the average model flow at density σ_θ = 25.5 psu for (a) the control run and (b) the lower sea level run. Strong gradients indicate high velocities. Black arrows indicate direction of positive transport. The black rectangle indicates the location of Figure 3.

velocity of the bottom 75 m, not the bottom velocity itself. The bottom layer velocity in the deepest portion of the channel is almost 12 cm s^{-1} at 27°N (Section 5) where the section across the straits is most constricted. If the model layers were thinner, presumably the bottom layer velocity would be smaller and the geostrophic transport calculated using boundary margin density and a reference velocity of zero at the deepest part of the section would be more accurate. Actual bottom velocities at the deepest part of the channel at 27°N are less than 10 cm s^{-1} [*Leaman et al.*, 1989], which would lead to a more modest discrepancy between the transport referenced to zero velocity and the true transport than the calculations using the model data.

3.4. Core Locations

[21] Of special interest are the core locations of *Lund et al.* [2006] and *Lynch-Stieglitz et al.* [2009], just west of the Dry Tortugas and off the Great Bahama Bank. Although these locations, chosen for their high Holocene sediment accumulation rates, are not exactly across the strait from each other, these high resolution cores have been used to assess flow variability over the Holocene and longer cores collected at the same locations will be used for a more precise constraint on the LGM transport through the Florida Straits.

[22] Geological constraints dictated that the vertical core profiles, while on either side of the Florida current, did not form a section perpendicular to the flow. In Figure 5, we show model potential density profiles at sections incorporating the core locations. We see that the vertical density profile at the Bahamas is very similar to the density profile at Cuba, on the same side of the Florida Current further upstream at the Dry Tortugas Section. When the density profile from the Bahamas at Section 3 (actual core loca-

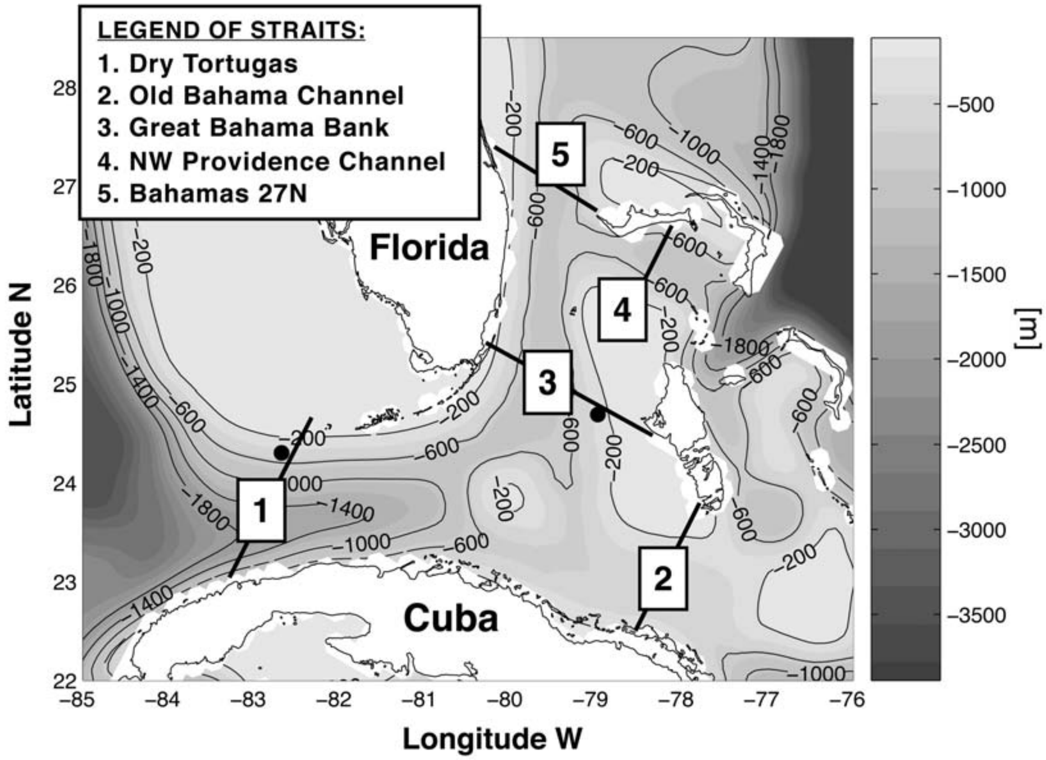

Figure 3. Florida Straits bathymetry and locations of sections across which transport in Table 1 is measured. The black dots indicate the *Lund et al.* [2006] and *Lynch-Stieglitz et al.* [2009] core locations.

tions) is substituted for the density profile from Cuba, the geostrophic flow relative to 850 m at Section 1 changes by only 0.7 Sv. This indicates that, at least in the model, the changes in flow structure between Section 1 and Section 3 are accommodated mostly by density changes along the Florida margin. This gives us increased confidence that the calculations from *Lund et al.* [2006] and *Lynch-Stieglitz et al.* [2009] are representative of the flow changes along Section 1 (Dry Tortugas Section).

[23] We next examine the choice of a reference level for a transport calculation across the Dry Tortugas Section. While *Lund et al.* [2006] assume a reference level at a depth of 850 m, the model velocity approaches zero closer to a depth of 1000 m (Figure 4) at the Dry Tortugas section. Geostrophic margin transport referenced to 850 m at the Dry Tortugas is 33.8 Sv, 4 Sv less than the 37.9 Sv referenced to the bottom. Observations across a 1050 m deep Florida Straits section show velocity approaching zero at 1000 m and a small (<0.05 m s^{-1}) velocity component at 850 m [*Leaman et al.*, 1995]. This small deviation from nonzero velocity at the 850 m reference level would correspond to an underestimate in geostrophic transport increase of about 2 Sv. Since the Great Bahama Bank section is about 850 m deep, any deeper transport at the Dry Tortugas section or barotropic transport due to bottom velocity at the Great Bahama Bank section, if there is any, will not be recorded

Table 1. Transport in Sverdrups for Observations, the Control Run, and the Lower Sea Level Run Across Sections Illustrated in Figure 3[a]

Section	Observed	Control Transport[b]				Lower Sea Level Total[c]
		Total	Geostrophic	Geostrophic Margins Calculation	Geostrophic Margins + Bottom v	
1. Dry Tortugas	28.4	39.5 ± 4.7	37.5 ± 5.9	37.9	37.8	37.1 ± 4.2
2. Old Bahama Channel	−1.9	2.2 ± 3.2	2.0 ± 2.9	1.3	2.1	4.2 ± 2.7
3. Great Bahama Bank	N/A	37.1 ± 2.1	40.3 ± 3.1	29.2	36.1	32.9 ± 2.1
4. NW Providence Channel	−1.2	−2.5 ± 1.3	−2.5 ± 1.0	−2.0	−2.1	−2.0 ± 1.0
5. Bahamas 27°N	31.5	39.8 ± 2.3	39.6 ± 2.3	31.0	40.0	34.8 ± 2.0

[a]Model error is 1 standard deviation.
[b]Run 1.
[c]Run 2.

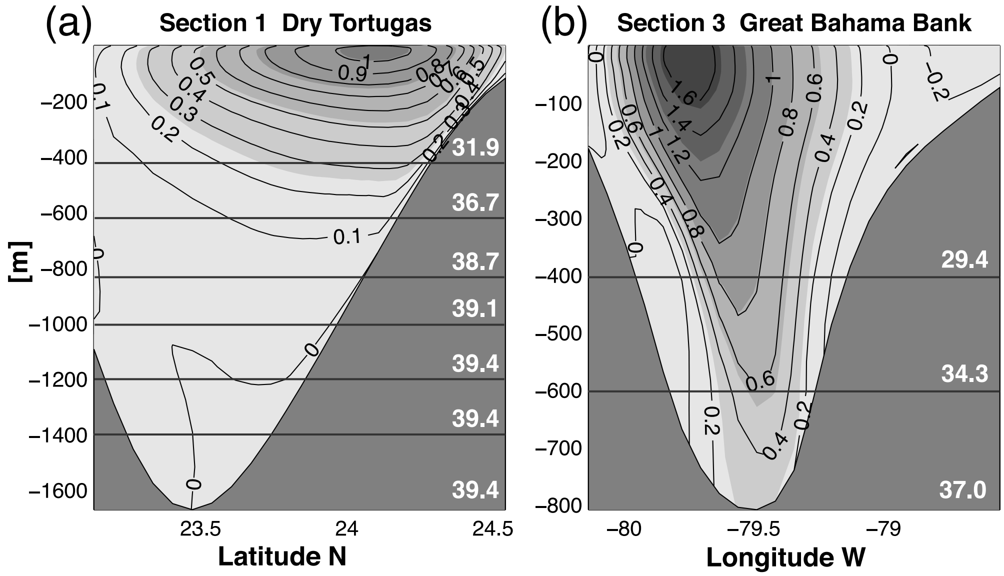

Figure 4. Velocity (m s^{-1}) profiles contoured across (a) the Dry Tortugas section (section 1 in Figure 3) and (b) the Great Bahama Bank section (section 3 in Figure 3). The numbers in each section indicate total transport (Sv = 10^6 m^3 s^{-1}) above the solid horizontal line reference levels.

by methods of the *Lund et al.* [2006] and *Lynch-Stieglitz et al.* [2009] study. In order to take into account any deeper flow in our future LGM geostrophic calculations, data points deeper than the Great Bahama Bank cores would have to be collected on the Cuban margin.

[24] While we have not investigated the ability of the core locations used in the *Lynch-Stieglitz et al.* [1999b] study to capture the geostrophic transport through the straits, the cores on the Florida margin were mostly downstream of the more constricted part of the Florida Straits. For these locations (e.g., 27°N section), the velocity in the bottom model layer is significant and when geostrophic transport is calculated referenced to the bottom of the channel transport is underestimated by 7 Sv. However, because the bottom layer is quite thick in the model (75 m at Sections 3 and 5), the error may not be quite as severe when the bottom layer velocity is neglected in a geostrophic estimate using sediment cores. However, while the flows in the deeper parts of the channel today are relatively weak, there is no guarantee that bottom velocities are small for the LGM. The caution voiced in the *Lynch-Stieglitz et al.* [1999b] paper that a larger barotropic component is possible for this northern section during the LGM still stands.

4. Conclusion

[25] Our results show that the effect of a 120 m sea level drop on the Florida Current transport is twofold: some of the water (2.3 Sv) flowing below 640 m at present is blocked by the sill and never reaches the Florida Straits, while some (2.5 Sv) shallows out and is diverted through the 500 m deep Old Bahama Channel (due to the Cay Sal Bank sill) with a reduced NW Providence Channel inflow (due to the 27°N sill) to rejoin the Gulf Stream north of the Florida Straits. The total flow deflection of 5 Sv is a maximum limit rather than an estimate, due to the larger than observed Florida Straits flow in the model (39.5 Sv). Out of the approximately 12–15 Sv geostrophic reduction inferred by *Lynch-Stieglitz et al.* [1999b], it is unlikely that more than 5 Sv could have been due to a sea level change.

[26] The geostrophic shear in the Florida Straits can be accurately reconstructed using ocean margin density (Table 1)

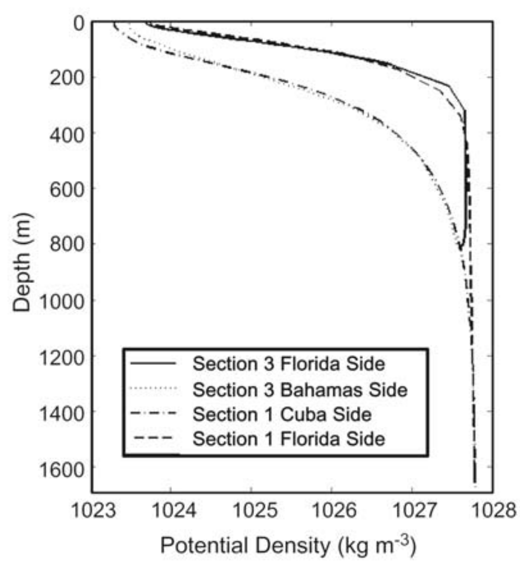

Figure 5. Potential density profiles (kg m^{-3}) from the straits margins across the Dry Tortugas section (section 1 in Figure 3) and the Great Bahama Bank section (section 3 in Figure 3) for the control run.

and the geostrophic transport can be accurately reconstructed when the bottom velocity is small (for the deeper Dry Tortugas section) or when the bottom layer velocity of the model is used to reference the calculation (the shallower sections). The deep Dry Tortugas section is the preferred location for paleoestimates of geostrophic transport, due to a safe assumption of zero bottom velocity.

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References

Alley, R. B., et al. (2005), Ice-sheet and sea-level changes, *Science*, *310*(5747), 456–460.

Bard, E., B. Hamelin, R. G. Fairbanks, and A. Zindler (1990), Calibration of the ^{14}C timescale over the past 30,000 years using mass spectrometric U-Th ages from Barbados corals, *Nature*, *345*, 405–410, doi:10.1038/345405a0.

Baringer, M. O., and J. C. Larsen (2001), Sixteen years of Florida Current transport at 27°N, *Geophys. Res. Lett.*, *28*(16), 3179–3182, doi:10.1029/2001GL013246.

Bassett, S. E., et al. (2005), Ice sheet and solid earth influences on far-field sea-level histories, *Science*, *309*(5736), 925–928.

Bulgakov, S. N., N. P. Bulgakov, and V. A. Ivanov (2003), A comparison of direct observations of velocity and transport in the Windward Passage, *Geofis. Int.*, *42*(1), 115–129.

Curry, W. B., and D. W. Oppo (2005), Glacial water mass geometry and the distribution of $\delta^{13}C$ of ΣCO_2 in the western Atlantic Ocean, *Paleoceanography*, *20*(1), PA1017, doi:10.1029/2004PA001021.

Di Lorenzo, E. (2003), Seasonal dynamics of the surface circulation in the Southern California Current System, *Deep Sea Res., Part II*, *50*, 2371–2388, doi:10.1016/S0967-0645(03)00125-5.

Haidvogel, D. B., et al. (2008), Ocean forecasting in terrain-following coordinates: Formulation and skill assessment of the Regional Ocean Modeling System, *J. Comput. Phys.*, *227*(7), 3595–3624.

Hamilton, P., J. C. Larsen, K. D. Leaman, T. N. Lee, and E. Waddell (2005), Transports through the Straits of Florida, *J. Phys. Oceanogr.*, *35*, 308–322, doi:10.1175/JPO-2688.1.

Johns, W. E., T. L. Townsend, D. M. Fratantoni, and W. D. Wilson (2002), On the Atlantic inflow to the Caribbean Sea, *Deep Sea Res., Part I*, *49*, 211–243, doi:10.1016/S0967-0637(01)00041-3.

Lambeck, K., and J. Chappell (2001), Sea level change through the last glacial cycle, *Science*, *292*(5517), 679–686.

Leaman, K. D., et al. (1989), The average distribution of volume transport and potential Vorticity with temperature at 3 sections across the Gulf-Stream, *J. Phys. Oceanogr.*, *19*(1), 36–51.

Leaman, K. D., P. S. Vertes, L. P. Atkinson, T. N. Lee, P. Hamilton, and E. Waddell (1995), Transport, potential vorticity, and current/temperature structure across Northwest Providence and Santaren Channels and the Florida Current off Cay Sal Bank, *J. Geophys. Res.*, *100*(C5), 8561–8569, doi:10.1029/94JC01436.

Lund, D. C., J. Lynch-Stieglitz, and W. B. Curry (2006), Gulf Stream density structure and transport during the past millenium, *Nature*, *444*, 601–604, doi:10.1038/nature05277.

Lynch-Stieglitz, J., et al. (1999a), A geostrophic transport estimate for the Florida Current from the oxygen isotope composition of benthic foraminifera, *Paleoceanography*, *14*(3), 360–373, doi:10.1029/1999PA900001.

Lynch-Stieglitz, J., et al. (1999b), Weaker Gulf Stream in the Florida Straits during the Last Glacial Maximum, *Nature*, *402*, 644–648, doi:10.1038/45204.

Lynch-Stieglitz, J., et al. (2007), Atlantic meridional circulation during the Last Glacial Maximum, *Science*, *316*, 66–69, doi:10.1126/science.1137127.

Lynch-Stieglitz, J., et al. (2009), Florida Straits density structure and transport over the last 8,000 years, *Paleoceanography*, *24*, PA3209, doi:10.1029/2008PA001717.

Malloy, R. J., and R. J. Hurley (1970), Geomorphology and geologic structure: Straits of Florida, *Geol. Soc. Am. Bull.*, *81*(7), 1947–1972.

Marchesiello, P., J. C. McWilliams, and A. Shchepetkin (2003), Equilibrium structure and dynamics of the California Current System, *J. Phys. Oceanogr.*, *33*, 753–783, doi:10.1175/1520-0485(2003)33<753:ESADOT>2.0.CO;2.

Milliff, R. F., and J. Morzel (2001), The global distribution of the time-average wind stress curl from NSCAT, *J. Atmos. Sci.*, *58*(2), 109–131, doi:10.1175/1520-0469(2001)058<0109:TGDOTT>2.0.CO;2.

Peltier, W. R., and R. G. Fairbanks (2006), Global glacial ice volume and Last Glacial Maximum duration from an extended Barbados sea level record, *Quat. Sci. Rev.*, *25*, 3322–3337.

Powell, B. S., et al. (2008), 4DVAR data assimilation in the Intra-Americas Sea with the Regional Ocean Modeling System (ROMS), *Ocean Modell.*, *25*(3-4), 173–188.

Richards, D. A., et al. (1994), Maximum sea levels for the Last Glacial Period from U-series ages of submerged speleothems, *Nature*, *367*(6461), 357–360.

Shchepetkin, A. F., and J. C. McWilliams (2005), The regional oceanic modeling system (ROMS): A split-explicit, free-surface, topography-following-coordinate oceanic model, *Ocean Modell.*, *9*(4), 347–404.

Smith, W. H. F., and D. T. Sandwell (1997), Global sea floor topography from satellite altimetry and ship depth soundings, *Science*, *277*, 1956–1962, doi:10.1126/science.277.5334.1956.

Toscano, M. A., and J. Lundberg (1999), Submerged Late Pleistocene reefs on the tectonically-stable SE Florida margin: High-precision geochronology, stratigraphy, resolution of substage 5a sea-level elevation, and orbital forcing, *Quat. Sci. Rev.*, *18*(6), 753–767.

Yokoyama, Y., K. Lambeck, P. De Deckker, P. Johnston, and L. Keith Fifield (2000), Timing of the Last Glacial Maximum from observed sea-level minima, *Nature*, *406*, 713–716, doi:10.1038/35021035.

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