In the North Atlantic Ocean, flow of North Atlantic Deep Water (NADW), and of its ancient counterpart Northern Component Water (NCW), across the Greenland-Scotland Ridge (GSR) is thought to have played an important role in ocean circulation. Over the last 60 Ma, the Iceland Plume has dynamically supported an area which encompasses the GSR. Consequently, bathymetry of the GSR has varied with time due to a combination of lithospheric plate cooling and fluctuations in the temperature and buoyancy within the underlying convecting mantle. Here, we reassess the importance of plate cooling and convective control on this northern gateway for NCW flow during the Neogene period, following Wright and Miller (1996). To tackle the problem, benthic foraminiferal isotope data sets have been assembled to examine δ^{13}C gradients between the three major deep water masses (i.e., Northern Component Water, Southern Ocean Water, and Pacific Ocean Water). Composite records are reported on an astronomical timescale, and a nonparametric curve-fitting technique is used to produce regional estimates of δ^{13}C for each water mass. Confidence bands were calculated, and error propagation techniques used to estimate %NCW and its uncertainty. Despite obvious reservations about using long-term variations of δ^{13}C from disparate analyses and settings, and despite considerable uncertainties in our understanding of ancient oceanic transport pathways, the variation of NCW through time is consistent with independent estimates of the temporal variation of dynamical support associated with the Iceland Plume. Prior to 12 Ma, δ^{13}C patterns overlap and %NCW cannot be isolated. Significant long-period variations are evident, which are consistent with previously published work. From 12 Ma, when lithospheric cooling probably caused the GSR to submerge completely, long-period δ^{13}C patterns diverge significantly and allow reasonable %NCW estimates to be made. Our most robust result is a dramatic increase in NCW overflow between 6 and 2 Ma when dynamical support generated by the Iceland Plume was weakest. Between 6 and 12 Ma a series of variations in NCW overflow have been resolved.

Components: 13,280 words, 12 figures, 3 tables.

Keywords: Northern Component Water; Greenland-Scotland Ridge; Iceland; plume; Neogene; δ^{13}C.
1. Introduction

[2] Wright and Miller [1996] argue that there is a strong negative correlation between the percentage of Northern Component Water (%NCW), which affects δ13C values in the Southern Ocean, and vertical motions of the Greenland-Scotland Ridge (GSR) during Neogene times (Figures 1–3). These vertical motions occur with at least two timescales. First, long-term subsidence of the GSR (timescale of 10s of millions of years) is controlled by cooling of the lithospheric plate as it drifts away from the mid-oceanic ridge. Secondly, short-term uplift and subsidence excursions (timescale of 1–10 Ma) are controlled by rapid buoyancy fluctuations within the Iceland Plume. This plume is a major convective upwelling within the mantle which has had an important effect upon the structural development of the North Atlantic Ocean over the last 60 Ma.

[3] Such changes in elevation of the GSR could affect flux of NCW into the Southern and Pacific Oceans. When the GSR is deeper (i.e., the Iceland Plume is cooler) a larger volume of deep water can flow over the ridge, thus increasing %NCW present in the Southern Ocean (Figure 3). When the Iceland Plume is hotter and the GSR is shallower, production of NCW could be significantly reduced. This lock-gate mechanism is important because it might help to improve our understanding of temporal changes within the Iceland Plume and thus contribute to knowledge of the dynamics of mantle convection. It may also help to gauge the strength of thermohaline circulation and hence the influence of variations in NCW production on Neogene climate.

[4] Present-day NCW (i.e., North Atlantic Deep Water, NADW) has three principal sources (Figure 1): (1) overspill from the Norwegian-Greenland Seas through the Faroe-Shetland Channel (1.7 Sv); (2) overspill through the Denmark Straits (2.9 Sv); and (3) convective overturning within the Labrador Sea [Dickson and Brown, 1994]. It remains unclear when the Labrador Sea became an important source of deep water [Mikolajewicz and Crowley, 1997; Burton et al., 1997; Lear et al., 2003]. Roughly 40% of the total NADW flux is sourced from the Labrador Sea [Schmitz and McCartney, 1993]. We assume that if the cold deep core of NADW from the Faroe-Shetland Channel and from the Denmark Straits is reduced significantly, then a large reduction in %NCW in the Southern Ocean results because the cold deep core of NADW consists primarily of GSR overflow water and not of warmer, less salty Labrador Sea Water.

[5] At present, flow of cold, deep NCW from the Norwegian-Greenland Sea into the North Atlantic Ocean is impeded by the GSR, which extends east and west from Iceland (Figure 1b). The present-day depth of the deeper channels through this ridge is no greater than ~850 m and allows a total of 5.6 Sv (10^6 m^3) of deep water to flow over the barrier [Dickson and Brown, 1994], although its volume increases by entrainment downstream of channel exits. Hence Norwegian Sea Overflow Water (NSOW) forms much of the Lower North Atlantic Deep Water (LNADW). At present, the core of NSOW flows at depths of 500–600 m, descending as it travels southward through the Faroe-Bank Channel [Kuijpers et al., 2002]. Cold northern waters are restricted to ~150 m above the seabed in this deepest gateway (Figure 1b). Therefore GSR overflow waters should be sensitive to plume-related dynamic uplift.

[6] Over the last 10 years, much new benthic foraminiferal δ13C (and δ18O) data have become available. These new data have an average sample interval of up to ~5 ka and facilitate improved core-to-core correlation (Figures 4–6) [Billups et al., 2004]. North Atlantic records include data from ODP Site 925 [Billups et al., 1997, 1998; H. Pfuhl, written communication, 2004]. Pacific Ocean records include analyses from ODP Site 846 (0–7.4 Ma [Mix and Shackleton, 1995; Shackleton et al., 1995b]). Most of the Southern Ocean data is from ODP Legs 177 and 185. The geological
timescale has also undergone revision thanks to the availability of astronomically tuned records for the Neogene period [Lourens et al., 2004]. With a greatly improved database and timescale, we can more accurately construct end-member records of carbon isotopes. Finally, we present a new strategy for fitting $\delta^{13}C$ records with smooth functions and confidence bands. In this way, we can estimate the variation of NCW overflow and its associated uncertainty through time and revisit the Wright and Miller [1996] hypothesis.

2. Benthic Carbon Isotope Data Sets

[7] A range of proxies is used to understand paleocirculation patterns, including $\delta^{13}C$ and trace metals (e.g., Neodymium, Cadmium). Following Wright and Miller [1996], we exploit $\delta^{13}C$ because

Figure 1. (a) Bathymetric map of North Atlantic Ocean. Principal features include Greenland-Scotland Ridge (GSR), Iceland-Faroe Ridge (IFR), Faroe-Shetland Channel (FSC), Rockall Trough (RT), Denmark Straits (DS), Norwegian-Greenland Sea (NGS), and Irminger Basin (IrB). V-shaped ridges (VSRs) are prominent to the southwest of Iceland (see Figure 2). Arrows show deep-water currents which overflow GSR. Fluxes are in Sverdrups (1 Sv = $1 \times 10^6$ m$^3$/s) beside respective arrow [Dickson and Brown, 1994]. A–A’ shows location of section. (b) Schematic section across Greenland-Scotland Ridge redrawn from Miller and Tucholke [1983] and Wright and Miller [1996]. Depths given for major spill-points of NCW over GSR.
samples can be easily assigned ages, records are of sufficiently high resolution, and there is now a wealth of published and unpublished analyses. $^{13}$C is a useful tracer of water masses because it reflects the nutrient content of deep water [Ruddiman, 2001], although care must be taken in choosing species that accurately record the isotopic composition of seawater. North Atlantic Deep Water is $^{13}$C rich because surface waters from which it forms have previously been depleted in both nutrients and $^{12}$C (manifested by a decrease of 1%) within the subtropical Atlantic Ocean. In contrast, deep water formed in the Weddell Sea contains more nutrients and less $^{13}$C because surface waters sink without being ventilated. $^{13}$C is not a conservative tracer. However, given a residence time of deep water in the deep Atlantic of $\leq 300$ yrs [Broecker and Peng, 1982], it is reasonable to model benthic $^{13}$C as a conservative tracer. Benthic $^{13}$C can then be used to quantify the mixing of two water masses at a particular site [Kroopnick, 1985; Oppo and Fairbanks, 1987]. One minor complication in understanding benthic $^{13}$C gradients is the existence of the “Mackensen effect” [Mackensen et al., 1993], which can lead to extremely low $^{13}$C values. However, this effect is probably smaller than the ventilation effect and as far as is possible we use epibenthic foraminifera such as Cibicidoides wuellerstorfi, which should circumvent this problem.

2.1. Benthic $^{13}$C End-Members

Benthic stable isotope data have been compiled from sediment cores located in the North Atlantic Ocean, in the Atlantic sector of the Southern Ocean and in the Pacific Ocean (Figures 4 and 5). Each region acts as an end-member in the mixing equation of Oppo and Fairbanks [1987] which is used to find an estimate of the percentage overflow.
The estimate of $\%\text{NCW}$, $\text{NCW}$, is given by

$$
\text{NCW} = \frac{\delta^{13}\text{C}_{\text{SO}} - \delta^{13}\text{C}_{\text{PO}}}{\delta^{13}\text{C}_{\text{NA}} - \delta^{13}\text{C}_{\text{PO}}} \times 100,
$$

where $\delta^{13}\text{C}_{\text{SO}}$, $\delta^{13}\text{C}_{\text{PO}}$ and $\delta^{13}\text{C}_{\text{NA}}$ are estimates of $\delta^{13}\text{C}$ for the Southern, Pacific, and North Atlantic Oceans. Note that in the text, we refer to $\%\text{NCW}$ rather than its estimate.

It is important to choose the geographical location of each data set carefully so that the true variation of each end-member is reflected in the composite record. North Atlantic records are taken from sites which are sufficiently far north and from depths at which benthic foraminifera are likely to be fully bathed by the North Atlantic end-member. Southern Ocean sites should ideally be located beneath the mixing zone of Upper Circumpolar Deep Water (UCDW) and NCW, a boundary that we assume remains fixed through time. This mixing zone is likely to have changed both its geographical position and its depth, but we lack data to constrain any movement during Neogene times.

Pacific Ocean sites should reflect Pacific Ocean deep water and therefore should not include records from restricted sub-basins such as ODP Site 677 [Mix and Shackleton, 1995; Oppo et al., 1995]. The Pacific Ocean can exhibit east-west $\delta^{13}\text{C}$ gradients which result from nonconservative $\delta^{13}\text{C}$ effects (i.e., water-mass aging caused by accumulation of nutrients [Kroopnick, 1985]). Pacific $\delta^{13}\text{C}$ gradients are problematic but difficult to circumvent, given the geographical distribution of reliable Neogene high-resolution data [Whitman and Berger, 1993].

Sources of data used for each oceanic end-member are given in Figure 5 and in Tables 1–3. $\delta^{13}\text{C}$ data are quoted to the PDB standard and species disequilibria are corrected using the species offsets of Graham et al. [1981], Shackleton and Hall [1984], Müller et al. [1991], and Katz et al. [2003]. Where multiple samples are used or where individual data points are obviously outliers, $\delta^{13}\text{C}$ values obtained from more unreliable species have been removed [Katz et al., 2003]. We have also removed data points for which both carbon and oxygen isotopes were greater than two estimated standard deviations from the regression curve. We subsequently refitted the data using our local linear kernel regression procedure. The distribution of standard deviations then better approximated a normal distribution, which is an assumption made during the fitting procedure.

Records used to construct Atlantic and Pacific end-members are mostly similar to those used by Wright and Miller [1996]. Our North Atlantic composite record now includes ODP Site 925 for part of the late Miocene and Early Pliocene.

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Figure 4. Map showing thermohaline conveyor belt [Rahmstorf, 2002]. Solid black lines, transport pathways of bottom water; dark gray lines, transport pathways of cold, deep water; light gray lines, transport pathways of warm, surface water; ellipses, locations of major downwelling zones; solid black circles, sites of DSDP and ODP cores used in this study.
Billups et al. [1998] have shown that Site 925, at 3042 m mostly reflects NADW which is dominant to depths of 3000 m on Ceara Rise during the Early Pliocene. Despite its southerly location we therefore think that Site 925 adequately records the variation in δ13C of NCW. In constructing the Pacific end-member, we have also replaced ODP Site 677 with ODP Site 846 and have supplemented DSDP Site 289 with data from DSDP Sites 296, 574A and ODP Sites 806, 846. Our most significant revision concerns the Southern Ocean end-member. Recent ODP cruises have recovered sediment cores which span the age interval of interest and which lie beneath the modern mixing zone of CDW and NADW [Billups, 2002]. New material from Leg 177 Sites 1088

Figure 5. Assemblage of δ13C isotopic records determined from benthic foraminifera. (a) North Atlantic Ocean. (b) Southern Ocean. (c) Pacific Ocean. Each record is color-coded according to site (see Tables 1–3 for further details). Box shows location of Figure 14.
Table 1. Sources of Data Used in North Atlantic $\delta^{13}$C End-Member

<table>
<thead>
<tr>
<th>Site</th>
<th>Isotope Source</th>
<th>Age Model Source</th>
<th>Interval, Ma</th>
<th>No. Points</th>
</tr>
</thead>
<tbody>
<tr>
<td>552A</td>
<td>Shackleton and Hall [1984]</td>
<td>Shackleton and Hall [1984]</td>
<td>0–3.446</td>
<td>604</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>2.810–2.872</td>
<td></td>
</tr>
<tr>
<td>607</td>
<td>Ruddiman et al. [1989]</td>
<td>Ruddiman et al. [1989]</td>
<td>0.4–0.451</td>
<td>119</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Clement and Robinson [1986]</td>
<td>0.522–0.567</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>0.750–0.810</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>1.251–1.648</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>5.0–6.240</td>
<td>259</td>
</tr>
</tbody>
</table>

Table 2. Sources of Data Used in Southern Ocean $\delta^{13}$C End-Member

<table>
<thead>
<tr>
<th>Site</th>
<th>Isotope Source</th>
<th>Age Model Source</th>
<th>Interval, Ma</th>
<th>No. Points</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Billups (unpub. data)</td>
<td>as above</td>
<td>0.667–25.374</td>
<td>27</td>
</tr>
<tr>
<td>1090</td>
<td>Billups et al. [2004]</td>
<td>Billups et al. [2004]</td>
<td>15.94–24.19</td>
<td>1532</td>
</tr>
</tbody>
</table>
combination of Leg 154 magnetic susceptibility and reflectance data together with $\delta^{18}O$ records from sites 704 and 747, which have reliable magnetostratigraphy and continuous sedimentation, were used as reference sequences to which other data sets could be correlated. The age model for 9–14 Ma does have large uncertainties compared with data outside this range but the $\delta^{13}C$ records are sparse, so correlation to future astronomically tuned benthic oxygen data is unlikely to alter the age models significantly. When depth-age constraints were added to the overall age model, resulting sedimentation rates were checked to ensure that sedimentation rates varied smoothly except at known hiatuses and unconformities.

2.2.1. Spectral Analysis

Spectral analysis of the corresponding benthic oxygen isotope data provides an useful check on the quality of age models applied to the carbon isotope data. The oxygen isotopic composition of sea water primarily reflects global ice volume [Imbrie et al., 1984], which responds to changes in the seasonal input of insolation at different latitudes. These changes are controlled by three orbital parameters of the Earth: eccentricity (100 and 400 ka periods), obliquity (41 ka) and precession (19 and 23 ka).

Both Southern and Pacific Oceans show that significant orbital frequencies are present, mostly as a result of the inclusion of astronomically tuned data sets. All oxygen data are coherent with the calculated insolation curve for higher orbital periodicities in both these oceans. Precession frequencies are only weakly coherent (below 80% significance level) suggesting that data are poorly tuned at these periodicities. This weakness is unsurprising given the caveats mentioned above, the temporal range of the data, and the coarser sampling intervals for the Miocene epoch compared with Plio-Pleistocene times. North Atlantic composite records have the weakest periodicities with peaks in power which are barely significant compared with background noise.

3. Statistical Analysis

There are significant errors in $\delta^{13}C$ measurements and in their ages, both of which propagate through the differences in equation (1) to affect our estimate of %NCW. The source of error in $\delta^{13}C$ is both measurement error, whose standard deviation ranges between 0.05 and 0.1%, and in vital effect species correction factors, which [Graham et al., 1981] quote as ranging between 0.12 and 0.46%.

In order to calculate the variation of %NCW with time, each composite $\delta^{13}C$ record is first fitted with a smooth function. There are numerous ways in which this fitting can be carried out. Here, we outline an approach which has two important features. First, we allow the bandwidth of the kernel estimator to vary through time. This flexibility ensures that large variations in sample distribution are accommodated. Secondly, confidence bands for best-fitting functions are constructed. We give a brief discussion of the techniques used and refer the interested reader to Samworth and Poore [2005] for further details.

3.1. Local Linear Kernel Regression

Our first task is to fit a smooth curve through each composite record. Gaussian filters are often used for this purpose [e.g., Mix and Shackleton, 1995; Wright and Miller, 1996; Zachos et al., 2001]. In the statistical literature, this locally weighted mean would be referred to as a local constant kernel estimator with a Gaussian kernel or weight function. The more sophisticated local linear kernel estimator, however, has the advantage of significantly reduced bias, particularly at the
Figure 6. Assemblage of $\delta^{18}$O records determined from benthic foraminifera. (a) Records used as reference model. Records from (b) North Atlantic, (c) Southern, and (d) Pacific Oceans which were used to refine age models.
boundaries of the data, or in regions of high sample resolution [Fan and Gijbels, 1996]. The local linear estimator for \((x_i, y_i)\) pairs is given by

\[
\hat{m}_h(x) = \frac{1}{n} \sum_{i=1}^{n} \frac{(x_2 - x_1)(x_i - x)}{s_{2x0} - s_1^2} y_i
\]

where

\[
s_r = \frac{1}{n} \sum_{i=1}^{n} (x_i - x)^r K_h(x_i - x), \quad r = 0, 1, 2,
\]

and the scaled kernel, \(K_h(\cdot) = h^{-1}K(\cdot/h)\), satisfies \(\int K_h = 1\).

[17] For both local constant and local linear estimators, the crucial choice is that of the bandwidth, \(h\), which controls the amount of smoothing: too small a value of \(h\) results in a spiky, undersmoothed curve; equally if \(h\) is too large then the data are oversmoothed leading to a loss of the signal.

[18] Oceanographic records are generally fitted using a Gaussian filter with a constant bandwidth. The main difficulty with this approach is that data distribution is inevitably highly nonuniform as a consequence of records being pooled from different cores, sampled at different resolutions, and because of gaps and unconformities. A bandwidth which is appropriate in a region of high resolution data may be many times smaller than that in a region of sparse data. We have sidestepped this issue by dividing the age range into blocks of approximately equal sample resolution; we then chose a separate bandwidth for each block. Since we are less interested in the rate of change of \(\delta^{13}C\), continuity of first and second derivatives is not of primary concern. Pacific, Southern and North Atlantic composite records were divided into 21, 16 and 12 blocks, respectively, and appropriate bandwidths were chosen for each block. Our regression method slides the chosen kernel through the age range of the composite record. At boundaries of blocks, part of the kernel cannot be used to weight \((x_i, y_i)\) pairs, which results in a large bias. To reduce this effect, we take the half-width of the kernel and extend the age range of each block by this width.

### 3.2. SIMEX Algorithm

[19] In addition to the error in the \(y\)-direction associated with measuring \(\delta^{13}C\) level at a given age, there is also an error in the \(x\)-direction associated with assigning an age to a particular sample. This error arises from several sources, including the assumption of a constant sedimentation rate between age-depth control points and the fact that the ages and composite depths of these control points are not precisely known. Temporal errors have often been ignored in previous studies. However, errors-in-variables techniques have been developed in statistical analysis for handling this problem.

[20] We have applied the Simulation Extrapolation (SIMEX) algorithm of Cook and Stefanski [1994] and Carroll et al. [1995] to all data. The results for the first block of the Pacific Ocean record are shown in Figure 7. The SIMEX algorithm analyses the effect of adding additional noise of known standard deviation to the age of each point and then extrapolates backward to estimate the effect of having no noise at all. Addition of this extra noise tends to blur peaks and troughs in the estimated curve. The final SIMEX estimate helps to clarify these features.

#### 3.2.1. Confidence Bands

[21] We have calculated 95% confidence bands for each block in order to estimate uncertainties of each composite \(\delta^{13}C\) record. Each confidence band is interpreted as a region which contains the true (i.e., unobserved) \(\delta^{13}C\) variation over the age range of the block in question with a probability of 0.95. Many techniques for the construction of confidence bands in nonparametric regression problems have been published in the statistical literature. We looked in detail at those proposed by Xia [1998] and at two constructions proposed by Claeskens and van Keilegom [2003]. We prefer the Xia [1998] confidence bands which are generally wider than those of the first construction of Claeskens and van Keilegom [2003]. Their second construction was found to be very sensitive to small changes in bandwidth for our data. The Xia [1998] bands behave in a more robust and conservative way which is suitable for our purposes (Figure 8). These bands also allow bias of the regression estimate at peaks and troughs to be estimated so that bands are not simply centered around the regression estimate. Unfortunately, there is currently no theoretical support for calculating confidence bands around SIMEX estimates. Therefore we only calculate Xia bands around the bias-corrected local linear regression estimate.

[22] Confidence bands for blocks with small amounts of data (usually blocks with a small age range) must be treated with caution because of undercoverage problems. In other words, the prob-
ability that a confidence band contains the true regression curve may be less than 95% if there is a limited amount of data (i.e., /C20 data points per block). Intervals that may be particularly susceptible to this problem lie between 6 and 7 Ma, and 12.5 and 13.5 Ma in the Pacific Ocean.

4. End-Member Regression Estimates

Figure 8 shows regression estimates for the three end-members over the last 23 Ma. The width of each 95% confidence band suggests that we can consider these composite records in two parts. Before ~12 Ma, all three composite records are effectively identical with the same pattern occurring in each ocean. After ~12 Ma, the three composite records diverge by up to about 1.5%.

In general terms, our results confirm well-known behavior. Circumpolar Deep Water (CPDW) represents a mixture of NADW and recirculated deep water from the Indian and Pacific Oceans. Consequently, CPDW δ13C values lie between those of NADW (characterized by high δ13C and low nutrients) and Indo-Pacific Deep Water (low δ13C and high nutrients). Without any NADW influence, the δ13C of CPDW should decrease to Indo-Pacific values, regardless of how the oceanic inventory of 12C or 13C might have changed [Charles and Fairbanks, 1992]. Thus the observation that Atlantic, Southern and Pacific Ocean δ13C end-member records coincide prior to ~12 Ma suggests that significant NCW production began at about 12 Ma. Detailed age calibration and error analysis strengthens earlier suggestions based on carbon isotopes that NCW production was weak during middle Miocene times (16–11.6 Ma) but became more important after about 12 Ma [Woodruff and Savin, 1989, 1991; Wright et al., 1992; Wright and Miller, 1993]. The magnitude of a δ13C gradient, or the lack of any such gradient, may also be due to lower whole ocean nutrients [e.g., Delaney and Boyle, 1987] and may be the cause of the similarities between the North Atlantic, Southern and Pacific Ocean composite records during early and middle Miocene times.

Changes in the global carbon reservoir are thought to cause long-period variations in δ13C [Miller and Fairbanks, 1985; Vincent and Berger, 1985; Shackleton, 1987]. These reservoir-related δ13C variations are most noticeable in the period 23–12 Ma. For example, the Monterey Excursion between ~17.5 Ma and 13.5 Ma is a well-known feature of Miocene benthic δ13C records which is thought to be caused by a period of increased organic carbon burial [Vincent and Berger, 1985]. δ13C also decreases between ~22 and 20 Ma, which may correlate with shallowing of the CCD identified by van Andel [1975] and van Andel et al. [1975]. Lyle [2003] attributes the shallowing of the CCD to warming and expansion of shallow seas due to a 50 m sea-level rise that led to increased sequestration of CaCO3.

There are also short (1–2 Ma) intervals at 19 and 17 Ma when one oceanic mass significantly diverges from another. These anomalous intervals typically involve large negative excursions of Pacific benthic δ13C, suggesting that there was intermittent production of a deep-water source, possibly located in the Antarctic or North Atlantic Oceans, depending on the direction of the δ13C gradient. North Atlantic type surface water that sinks to form deep water is enriched in 13C, but as deep waters move away from their source region they gradually accumulate organic carbon rich in 12C from shal-
Figure 8. Regression estimates for three sets of composite records. In each case, other two regression estimates are shown in pale shading for comparison. (a) North Atlantic Ocean. Solid black circles, observations; solid black curve, regression estimate; bands of different red shades, 95% confidence bands for each block of data. (b) Southern Ocean. As above with green shades. (c) Pacific Ocean. As above with different blue shades. Note overlap in confidence bands between 23 and 12 Ma.
low waters. Thus its $\delta^{13}C$ signature is gradually reduced as it travels from its source (i.e., water-mass aging).

[27] From 12–0 Ma, rapid divergence of the end-member curves may mark the beginning of more sustained large volume production of NCW. The North Atlantic end-member oscillates around a constant value from this time onward, reflecting continuous sinking of water depleted in $^{12}C$ by surface productivity. The Pacific Ocean end-member has the lowest values of $\delta^{13}C$ during this period. The Southern Ocean end-member fluctuates between North Atlantic and Pacific Ocean values. For the purposes of this paper, we use $\delta^{13}C$ as a water-mass tracer and assume that changes in Southern Ocean $\delta^{13}C$ are caused by variations in the strength of NCW export from the Norwegian Sea. These variations can only give an indication of the relative flux of NCW in the Southern Ocean and should not be regarded as a measure of absolute export fluxes.

4.1. Estimation of Northern Component Water 0–23 Ma

[28] Following Wright and Miller [1996], we wish to use these records to estimate the variation in NCW overflow over the last 23 Ma. According to Oppo and Fairbanks [1987], the percentage of NCW present in the Southern Ocean is given by equation (1). This mixing equation is undoubtedly a simplistic way of representing the relationship between these three water masses, but it can be regarded as a useful starting point. It can only be applied when $\delta^{13}C_{SO} \geq \delta^{13}C_{PO}$ and $\delta^{13}C_{NA} > \delta^{13}C_{PO}$. Thus before 12 Ma, when all three confidence bands overlap, we cannot calculate %NCW [contra Wright and Miller, 1996].

[29] Between 12 and 0 Ma, %NCW can be calculated (NCW) and an estimate of its variance is given by

$$\tilde{\nu}_{NCW} = \frac{A}{(\delta^{13}C_{NA} - \delta^{13}C_{PO})^2} + \frac{B}{(\delta^{13}C_{NA} - \delta^{13}C_{PO})^2} - \frac{C}{(\delta^{13}C_{NA} - \delta^{13}C_{PO})^2},$$

(3)

where $A = \tilde{\nu}^{SO} + \tilde{\nu}^{PO}$, $B = (\delta^{13}C_{SO} - \delta^{13}C_{PO})(\tilde{\nu}^{NA} + \tilde{\nu}^{PO})$ and $C = (\delta^{13}C_{SO} - \delta^{13}C_{PO})\tilde{\nu}^{PO}$ (see Serfling [1980] for further details). Here, $\tilde{\nu}^{SO}$, $\tilde{\nu}^{PO}$ and $\tilde{\nu}^{NA}$ are the estimated variances of $\delta^{13}C_{SO}$, $\delta^{13}C_{PO}$ and $\delta^{13}C_{NA}$, respectively.

[30] %NCW is shown in Figures 9 and 10. We must emphasize that %NCW is unlikely to decrease to zero at any time during the last 12 Ma and so the percentage range is best thought of as indicating “weaker” and “stronger” NCW overflow. When %NCW is close to zero, it may partly indicate movement of the mixing zone northward or shoaling of the mixing zone such that Southern Ocean sites no longer sit within this zone. During the last 12 Ma, the most striking feature of calculated %NCW is the rapid increase at 6 Ma and decrease at 2 Ma. Other short period variations also appear to be well-resolved.

4.2. Evolutive Spectral Analysis

[31] The composite records span a large interval and we expect that amplitudes of different frequency responses to insolation will vary with time. These records have been carefully constructed and are of high sample resolution so it is useful to study changes in power spectra of different orbital periodicities contained within the data. An evolutive spectral analysis technique is used to display temporal changes in the power of different orbital periodicities. To apply this technique, results from the local linear kernel regression are exploited because bandwidths were chosen to reflect sample resolution, enabling short-period cyclicities, where present, to be retained. In some age intervals, there is limited spectral information for orbital frequencies of interest. Evolutive spectra are generated using the Thomson multi-taper method with a time-bandwidth product of 4 and eigenspectra of 6 [Torrence and Compo, 1998]. Nonparametric regression estimates were analyzed with a sliding window length of 1500 ka and a step size of 125 ka. The shorter %NCW record was analyzed with a window length of 625 ka and a step size of 62.5 ka.

[32] Individual records show that, where there is significant data resolution, nonparametric analysis of composite records retains the orbital cyclicities expected from benthic stable isotope records. We have also examined the spectral content of the 0–7 Ma %NCW estimate since there are spectral peaks in the benthic $\delta^{13}C$ data whence it is calculated (Figure 10). The presence of obliquity is notable between 3 and 1.6 Ma, although it is present quite consistently across the entire 0–7 Ma interval (Figure 11). This strong obliquity signal suggests that %NCW is sensitive to obliquity forcing. Such a response may be characteristic of either the way carbon cycles respond to insolation forcing in the oceans during Late Neogene
times, or a determinant of NCW formation in the Northern Seas. It is interesting to note that the obliquity component of the North Atlantic end-member during this interval decays from a peak around 6.5 Ma and is weak by 5 Ma. In contrast, the Pacific Ocean benthic δ^{13}C record only contains a large obliquity component from ~3.5 Ma. [Haug et al., 2005] suggest that changes in the subarctic Pacific halocline, triggered by an obliquity minimum at ~2.7 Ma, resulted in moisture release and in the Northern Hemisphere Glaciation (NHG). The %NCW record may be responding to obliquity forcing in the Pacific Ocean beginning around 3 Ma. The timing of changes in obliquity amplitude in the Pacific benthic δ^{13}C record occur before the changes in the surface subarctic Pacific Ocean. The increase in obliquity component in %NCW ~3 Ma, however, does coincide with Northern Hemisphere Glaciation. The Pleistocene benthic δ^{13}C stack of Lisiecki and Raymo [2005] also shows changes in orbital forcing response, which occurred in conjunction with the climate transition ~3 Ma. Their data indicate a change in the phase (or lag) of the precession and obliquity components of benthic δ^{18}O with respect to an orbitally forced ice-volume model at this time. Haug and Tiedemann [1998] suggest that NHG was caused by a combination of the late Cenozoic cooling trend, Panama Isthmus closure affecting deep-water circulation in the North Atlantic, and finally triggered by favorable obliquity forcing. Recent modelling to test this hypothesis has failed to find support that the delivery of moisture (through closure of the Panama Isthmus) was a precondition for the obliquity trigger [Klocker et al., 2005]. The timing of the impact of closure of the Central American Seaway on deep-water circulation ~4.6 Ma [Ravelo et al., 2004] also casts doubt on the closure of the Panama Isthmus as the final precondition for obliquity-triggered NHG [Haug et al., 2005]. Klocker et al. [2005] suggest that whilst obliquity modulation may be the trigger, the climate system needed different preconditions to allow the long-term buildup of ice in the Northern Hemisphere.

5. Iceland Plume Variation?

[Wright and Miller [1996] suggested that the variation of NCW flux was controlled by vertical motions of the GSR which were in turn controlled by temperature fluctuations within the Iceland Plume. Here we wish to test this hypothesis in the light of our newly calculated %NCW and our study of the associated errors (Figures 9 and 10).

5.1. Gateway Closure

[33] Wright and Miller [1996] suggested that the variation of NCW flux was controlled by vertical motions of the GSR which were in turn controlled by temperature fluctuations within the Iceland Plume. Here we wish to test this hypothesis in the light of our newly calculated %NCW and our study of the associated errors (Figures 9 and 10).
ways. We are also conscious of recent suggestions that orbital forcing may combine with gradual changes in atmospheric and circulation boundary conditions to produce large, long-term changes in parameters such as benthic $\delta^{18}$O and %NCW [e.g., Haug and Tiedemann, 1998; Haug et al., 2005]. Here, we restrict ourselves to tectonic explanations, although we acknowledge that there may be other (contributory) causes of %NCW variations.

[35] Two crucial issues affecting NCW production are convective overturning within the Labrador Sea and changes in inter-oceanic communication. The Labrador Sea is evidently an important source of deep water whose temporal evolution is poorly understood. For now, we must assume that its contribution does not vary significantly over the interval of interest. On the other hand, opening and closing of gateways could profoundly influence thermohaline circulation (including Labrador Sea convective overturn) and we must address the gateway issue in more detail.

[36] Two significant gateway closures occur during the interval of interest: progressive closure of the Panama Gateway and final closure of the Tethyan Ocean. In both cases, deep-water circulation patterns and $\delta^{13}$C gradients between the major oceanic basins were undoubtedly affected. Tethyan ocean closure is imprecisely dated but it probably influenced Atlantic deep waters until ~13 Ma [Ramsay et al., 1998].

[37] Onset of significant NCW production may have been influenced by closure of the Panama Gateway, which gradually separated Atlantic and Pacific oceanic circulation through Miocene-Pliocene times. Global circulation models have been used to suggest that the present-day configuration of global thermohaline circulation is not possible with an open Panama gateway [Maier-Reimer et al., 1990]. If the pathway of warm, salty water associated with the Gulf Stream is forced to the north, it is possible that the Labrador Sea becomes the site of important deep-water production. It is also likely that, whilst deep-water production was possible in
the Norwegian–Greenland Seas prior to the formation of the Panama Isthmus, the flux was significantly increased by the Gulf Stream bringing warm salty surface waters to high latitudes. Both changes could significantly influence the quantity of NCW exported into the Southern Ocean. Our principal consideration here is timing of gateway closure. Foraminiferal stratigraphy of Central America suggests that initial uplift of the Panama Sill occurred at 15–15.9 Ma, although the deep-water connection was probably not severed at this time [Duque-Caro, 1990]. Separation of Pacific and Caribbean benthic faunas, following a hiatus at 12.2–13.1 Ma, represents the first evidence for development of a sill which was shallow enough to restrict Atlantic-Pacific deep water-mass exchange [Duque-Caro, 1990]. Nesbitt and Young [1997] also argue that closure began at about 12 Ma. Some deep water-mass exchange may have continued until 6.7 Ma [McDougall, 1996]. Studies of planktonic faunas indicate gradual separation of surface-water flow after this time, suggesting complete divergence of surface-dwelling faunas by 1.8 Ma [Keller et al., 1989]. Others have argued that water depths of less than 100 m were reached by 4.6 Ma on the basis of surface salinity increases recorded by planktonic foraminifera [Keigwin, 1982; Haug and Tiedemann, 1998].

[38] In summary, gradual restriction of deep-water flow relatively early in the history of closure of the Panama gateway could have played a role in affecting major NCW production during mid-late Miocene times. Although there is some disagreement about the exact timings of shallowing and closure, the consensus in the literature suggests that shoaling to 100 m occurred at about 4.6 Ma [Haug and Tiedemann, 1998] and that complete closure occurred at about 3.5 Ma [Duque-Caro, 1990; Haug and Tiedemann, 1998; Frank et al., 1999] (Figure 10). Neither event appears to correspond with major changes in the δ¹³C composite records or %NCW.

5.2. Long-Term Subsidence of Greenland-Scotland Ridge

[39] Long-term subsidence of the GSR in response to plate cooling could have exerted a primary control on onset of NCW production [Vogt, 1972]. Subsidence histories of the three present-day GSR gateways are considered separately since they have different tectonic histories (Figure 1). The Faroe-Shetland Channel is floored by a conti-
We suggest that subsidence of the Greenland-Scotland Ridge on either side of Iceland exerted a direct control on NCW formation that was as least as significant as closure of the Panama Gateway. At present, more than half of the northward flow of saline surface water travels across the Iceland-Faroe Ridge and more than half of the southward return flow of NADW uses the Denmark Straits. The Iceland-Faroe Ridge, and possibly also the Denmark Straits, probably lay above sea-level prior to middle Miocene times. Limited connection between the Atlantic Ocean and the Greenland-Iceland-Norway Sea served to inhibit formation of NCW north of Iceland. By about 12 Ma, it is likely that water depths in excess of 100 m had been established above the GSR immediately either side of Iceland, which would have allowed saline Atlantic surface water to flow northward, cool, and then sink to form NCW. Initially, return flow of NCW probably occurred through the Faroe-Shetland Channel, which has probably been open and relatively deep since Early Oligocene times [Davies et al., 2001]. Ongoing subsidence of oceanic crust around Iceland through late Miocene time allowed an increasing rate of NCW export through the Denmark Straits, which may account for coeval divergence of the Atlantic and Pacific δ13C records.

5.3. Short-Term Uplift/Subsidence of Greenland-Scotland Ridge

Short-lived cycles of uplift and subsidence are superimposed upon the long-term subsidence of the GSR. Wright and Miller [1996] used a set of seismic reflection profiles that traverse the Reykjanes Ridge spreading axis to constrain the timing and amplitude of these cycles (Figures 1–3). Variations in depth to oceanic basement define a set of V-shaped ridges that are symmetric about the spreading axis and converge to the south [Vogt, 1971]. A coeval set of V-shaped ridges that converge to the north occur along the Kolbeinsey Ridge, north of Iceland (Figure 1), [Jones et al., 2002a]. These V-shaped bathymetric ridges reflect crustal thickness variations that are generated by thermal (and possibly compositional) anomalies within the head of the Iceland Plume that spread outward from Iceland. Fronts of relatively high temperature, propagating outward from Iceland, are associated with dynamic uplift of the overlying plate, since hotter regions are more buoyant. The geometry of the V-shaped ridges alone cannot be used to determine whether outward asthenosphere flow was radial or channelled beneath the spread-
ing axis [White and Lovell, 1997; Jones et al., 2002a]. It is more likely that fronts of high temperature propagated radially beneath a horizontal boundary at a depth of about 100 km where a large viscosity increase arises from dewatering of the mantle in the earliest stages of melting [Ito et al., 1999; Ito, 2001; Jones et al., 2002a]. Hence V-shaped ridges formed at the Reykjanes Ridge south of Iceland can be correlated with uplift of the GSR to the west and east of Iceland.

[42] V-shaped ridge bathymetry can be converted into residual depth, which is the difference between the observed depth to oceanic basement and the depth predicted by a half-space cooling model. Depths to oceanic basement are derived from Talwani et al. [1971], reported by Wright and Miller [1996]. Residual depth directly reflects dynamic support, although features less than about 0.1 Ma in duration may have been modified by faulting and can be regarded as noise. Wright and Miller [1996] also presented estimates of accumulation rate of major North Atlantic sediment drifts and estimates of lava extrusion rates at Iceland as supporting evidence that vertical motion of the GSR affected the flow of NCW. Jones et al. [2002a] produced a V-shaped ridge record based on gravity anomalies. The gravity record resembles a smoother version of the residual depth record.

[43] The V-shaped records from residual depth and gravity were originally plotted as functions of time using magnetic anomalies correlated to the timescales of Berggren et al. [1985] and Cande and Kent [1995], respectively [Wright and Miller, 1996; Jones et al., 2002a]. We have placed these records on the astronomical timescale of Lourens et al. [2004], since we wish to compare them with our %NCW record.

5.4. Timing of Uplift at the Major Spill-Points

[44] It is necessary to adjust the V-shaped ridge records to reflect the timing of uplift/subsidence events at locations of deep-water flow over the GSR in order to make a comparison with %NCW. Temperature anomalies that generate the V-shaped ridges propagate radially outward from Iceland, so the timing of associated uplift depends on the propagation speed and radial distance from the plume center. The propagation speed is determined from the angles between V-shaped ridges and the spreading axis. Wright and Miller [1996] measured propagation speeds between 1140 km/Ma for the youngest ridge and 163 km/Ma for their oldest ridge.

[45] The residual depth record was formed at 515 km from the plume center. Wright and Miller [1996] adjusted the original record to reflect uplift/subsidence at Iceland itself using a different propagation speed for each ridge. We removed these adjustments to recreate the original record. We then sought a uniform time shift that resulted in a good negative correlation between rapid changes in the slope of residual depth and %NCW. A uniform time shift of +0.11 Ma results in a reasonable negative correlation. Denmark Straits and the Faroe Shetland Channel currently lie 480 km and 760 km from the plume center, respectively. Using the range of propagation speeds determined by Wright and Miller [1996], a time shift of between +0.03 and +0.21 Ma applied to the residual depth record should predict uplift/subsidence of the Denmark Straits, and a shift of between −0.22 and −1.52 Ma should predict uplift/subsidence of the Faroe Shetland Channel. The velocities found by Wright and Miller [1996] progressively increase for each ridge and so the time shift required for the plume anomaly to reach Denmark Straits should be closer to +0.03 Ma for the younger ridges (Ridge 1) and closer to 0.21 Ma for the older ridges (Ridges 3 and 4). The residual depth record has been shifted by +0.11 Ma (Figure 9c) and by +0.21 Ma (Figure 9d) to reflect vertical motions of the Denmark Straits. We have also tried time shifts which vary with age, as well as automated correlation. In our view, the use of an elastic correction is not yet justified.

[46] Since a reasonable correlation between residual depth and %NCW requires a uniform time shift of +0.11, we suggest that vertical motions of the Denmark Straits have exerted the most important control on flow of deep water between the Norwegian-Greenland Sea and the Atlantic Ocean during the Neogene period. Comparing the V-shaped ridge gravity record with %NCW supports this conclusion, although the negative correlation is not so striking because the gravity record is smoother than the residual depth record. Note that Wright and Miller [1996] and Jones et al. [2002a] found that %NCW was sensitive to uplift at Iceland itself. The improved quality of our %NCW record suggests instead that NCW flow is more sensitive to uplift at the Denmark Straits.

6. Discussion

[47] Having made appropriate adjustments for the time taken for a hot pulse to reach a distall spill
point on the GSR (i.e., the Denmark Straits), we can now compare our revised %NCW record with uplift/subsidence records of the Iceland Plume. Plume-related uplift and plate cooling correlate with important changes in %NCW although we acknowledge that other events probably impact %NCW production and export during the Neogene period.

6.1. %NCW Between 23 and 12 Ma

Fluctuations in %NCW over the interval 23–12 Ma are inconclusive because the three composite $\delta^{13}$C records overlap within error. There is evidence for small inter-oceanic $\delta^{13}$C gradients but carbon isotopes do not necessarily reveal the presence of large-scale meridional circulation. Higher resolution data may help to reduce the width of confidence bands and to increase the precision of correlation between end-members, but it is unlikely that large $\delta^{13}$C gradients will be uncovered. The divergence of North Atlantic and Pacific Ocean $\delta^{13}$C records at 12 Ma is attributed to the production of large volumes of NCW in the North Atlantic/Norwegian Greenland Sea. Deep-water formation patterns are more easily understood by examining statistically significant %NCW variations.

6.2. Late Miocene to Present %NCW Variations: One Simple Cause?

Since ~12 Ma, large amounts of deep water have formed episodically in the North Atlantic Ocean, causing $\delta^{13}$C records for the North Atlantic and Pacific Oceans to diverge (Figure 8). Southern Ocean data fluctuate between these two end-members. Between 11 and 6 Ma, estimates of %NCW are reliable and negatively correlate with vertical motions of the GSR (Figure 9). This correlation is particularly striking when the residual depth record is shifted to reflect uplift and subsidence of the Denmark Straits. Thus plume-related uplift of Iceland appears to significantly restrict the amount of deep water which spills over the GSR at the Denmark Straits. Its impact would be greater than simply reducing the 5.6 Sv of water which currently overflows the ridge, since overflowed water entrains intermediate waters which help to increase the flux of present-day Lower North Atlantic Deep Water [Dickson and Brown, 1994]. A weaker correlation between this %NCW record and subsidence of the Faroe Shetland Channel suggests that the Denmark Straits controlled large-volume NCW production.

Figure 10 shows %NCW for the last 12 Ma when the highest resolution records exist for each of the three end-members. Several broad-scale features are of interest. At 6 Ma, there was a rapid increase in %NCW, which we interpret as an increase in NCW production rather than a proportional increase in Southern Component Water. Another rapid increase occurred at 5.2 Ma. These changes precede a sustained interval of NCW production which continues until 2.7 Ma. We attribute the increase in %NCW at 6 Ma to reduced dynamic support at the Denmark Straits. This reduction is inferred from the trough between Ridges 1 and 2. Wold [1994] has highlighted a dramatic increase in sedimentary flux across the GSR at ~6 Ma, and a slight decrease at ~2 Ma, both of which correlate very well with an increase and decrease in %NCW, respectively.

From 2.7 Ma onward, the proportion of NCW in the Southern Ocean diminishes steadily until ~1.9 Ma. Intensification of the Northern Hemisphere Glaciation and stagnation of waters in the northern seas also occurred at this time [Maslin et al., 1998]. This time interval coincides with the formation of V-shaped Ridge 1 in response to an increase in temperature within the Iceland Plume head. We suggest that the associated dynamic uplift of the GSR decreased southward overflow of deep water from the NGS, and reduced strength of thermohaline circulation. Stagnation of bottom waters and poor carbonate preservation reported by Henrich et al. [2002] ensued. Reduction in NCW also reduced latitudinal heat transport, thereby rapidly increasing the areal extent of Northern Hemisphere Glaciation.

The composite benthic $\delta^{18}$O record suggests that, despite large fluctuations in the proportion of NCW in the Southern Ocean, periods of sudden increase or decrease in %NCW are not generally associated with significant changes in deep-water $\delta^{18}$O composition. Thus long-term changes in the strength of NCW do not obviously correlate with major changes in the Earth’s glaciation history. However, two of the major decreases in benthic $\delta^{18}$O, at ~15 and ~3.6 Ma, do match the rapid growth of the two biggest V-shaped ridges formed during the Neogene period (Ridges 1 and 3). Our records suggest that changes in %NCW, caused by the formation of Ridge 1, are associated with an increase in Northern Hemisphere Glaciation at ~3.6 Ma [Wright and Miller, 1996]. The possibility remains, given the uncertainties associated with %NCW prior to ~12 Ma, that Ridge 3
interrupted NCW production and export from the Norwegian-Greenland Seas, providing moisture for the mid-Miocene growth in the Antarctic ice sheet [Schnitker, 1980].

6.3. Sensitivity of %NCW to Data Choice

Although %NCW generally correlates well with Iceland Plume activity, there was initially a significant mismatch at 7 Ma (Figure 9). When composite records are scrutinized, it is clear that the 7 Ma peak in %NCW is caused by an increase in the Southern Ocean $\delta^{13}C$ composite record that is controlled by a small number of high values (Figure 8). These high values of $\delta^{13}C$ come from two sets of analyses carried out on ODP site 704 at 2.5 km depth (Figure 12). These analyses diverge from the trends in ODP site 1088 and DSDP site 360 (at depths of 2 and 3 km, respectively). There is further disagreement between the two Site 704 data sets, Müller et al. [1991] and Wright et al. [1991], where Müller's data is generally positively offset from the other data. Billups et al. [2002] states that there is no obvious environmental cause for the differences between the two Site 704 data sets that sample the late Miocene period. We therefore suggest that data from Site 1088 (and Site 360) better reflect the benthic $\delta^{13}C$ composition of the Southern Ocean through this interval. If Site 704 values are removed, causing the record to be based on Site 1088 during the late Miocene, the %NCW peak at 7 Ma is considerably diminished (Figure 9). One cause of the difference in $\delta^{13}C$ between Site 704 and Sites 360 and 1088 may be small changes in the geometry of the mixing front between NADW and CDW [Billups et al., 2002]. The differences between Site 704 and Sites 360 and 1088 highlight the importance of choosing appropriate data for each of the benthic $\delta^{13}C$ end-members, since %NCW is sensitive to changes in data, especially in the Southern Ocean. Close examination of the records reveals no other region of overlap between data sets with such a large discrepancy.

7. Conclusions

In order to test Wright and Miller’s [1996] assertion that temporal variations associated with the Iceland Plume controlled overflow of Northern Component Water at the Greenland Iceland Ridge, we have assembled composite $\delta^{13}C$ records for the Atlantic, Southern and Pacific Oceans. These composite records were calibrated with the astronomical timescale and fitted with nonparametric functions. We paid particular attention to the construction of confidence bands.

Despite the many caveats concerning the use of $\delta^{13}C$ records to reconstruct overflow of NCW, and despite uncertainties surrounding the importance of Labrador Sea convective overturn and the timing of Panama Gateway closure, there is a
remarkably good correlation between %NCW and inferred vertical motions of the Denmark Straits. This correlation suggests that short-term (order 2–3 Ma) fluctuations of %NCW after 12 Ma are controlled by vertical motions of the GSR, which are generated by temperature fluctuations associated with the Iceland Plume, vindicating the Wright and Miller hypothesis. Uplift fluctuations are likely to be several hundred meters but detailed modelling is required to confirm this influence. The match between short-term features of the %NCW record and the V-shaped ridge record strengthens the suggestion that the Greenland-Scotland Ridge also controlled onset of NCW production in the longer term. Nonetheless, more sustained production of NCW may only occur after closure of the Panama Isthmus in order to supply warm salty surface water to high northern latitudes.

[57] Intensification of Northern Hemisphere Glaciation at 2.7 Ma may be related to the restriction in %NCW in response to dynamic uplift associated with the youngest V-shaped ridge. A reduction in %NCW at 2.7 Ma may coincide with an increase in obliquity forcing at 2.8 Ma. This reduction may indicate increased dominance of NCW, which has a strong obliquity component, in controlling circulation and glaciation. These hypotheses could, in part, be tested by detailed examination of North Atlantic ODP sites. Finally, the statistical techniques used to generate confidence bands for nonuniformly sampled isotopic data may be of more general application in paleoceanography.

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