Southeastern Atlantic deep-water evolution during the late-middle Eocene to earliest Oligocene (Ocean Drilling Program Site 1263 and Deep Sea Drilling Project Site 366)

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ABSTRACT

Comparison of new benthic foraminiferal δ18O and δ13C records from Ocean Drilling Program (ODP) Site 1263 (Walvis Ridge, southeast Atlantic, 2100 m paleodepth) and Deep Sea Drilling Project (DSDP) Site 366 (Sierra Leone Rise, eastern equatorial Atlantic, 2200–2800 m paleodepth) with published data from Atlantic and Southern Ocean sites provides the means to reconstruct the development of deep-water circulation in the southeastern Atlantic from the late-middle Eocene to the earliest Oligocene. Our comparison shows that in the late-middle Eocene (ca. 40 Ma), the South Atlantic was characterized by a homogeneous thermal structure. Thermal differentiation began ca. 38 Ma. By 37.6 Ma, Site 1263 was dominated by Southern Component Water; at the same time, warm saline deep water filled the deeper South Atlantic (recorded at southwest Atlantic ODP Site 699, paleodepth 3400 m, and southeast Atlantic ODP Site 1090, paleodepth 3200 m). The deep-water source to eastern equatorial Site 366 transitioned to Northern Component Water ca. 35.6–35 Ma. Progressive cooling at Site 1263 during the middle to late Eocene and deep-water thermal stratification in the South Atlantic may be attributed at least in part to the gradual deepening and strengthening of the proto–Antarctic Circumpolar Current from the late-middle Eocene to the earliest Oligocene, as the Drake and Tasman gateways opened. Our isotopic comparisons across depth and latitude provide evidence of the development of deep-water circulation similar to modern-day Atlantic Meridional Overturning Circulation.

INTRODUCTION

The late-middle Eocene to earliest Oligocene (ca. 38–33 Ma) was a period of transition to large-scale glaciation on Antarctica and cooling of the oceans (e.g., Miller and Fairbanks, 1985; Barron et al., 1991; Ehrmann and Mackensen, 1992; Miller et al., 1992, 2005a, 2008b, 2009; Diester-Haass and Zahn, 1996; Zachos et al., 1996; Lear et al., 2004, 2008; Coxall et al., 2005; Katz et al., 2008; Wade and Pearson, 2008; Pusz et al., 2011; Wade et al., 2012; Bijl et al., 2013). The transition from warmer to colder high-latitude ocean temperatures from the early Eocene to the Oligocene (ca. 50–33 Ma) is evident in many paleoceanographic records (e.g., Browning et al., 1996; Diester-Haass et al., 1996; see compilations of Cramer et al., 2009, 2011). Southern Ocean deep-water temperatures peaked during the early Eocene and subsequently cooled during the middle Eocene to the early Oligocene by as much as 5–10 °C (e.g., Mackensen and Ehrmann, 1992; Zachos et al., 2001; Diekmann et al., 2004; Cramer et al., 2011; Bohaty et al., 2012). The earliest documented ephemeral Antarctic glaciation occurred in the late-middle Eocene (ca. 37.3 Ma; Scher et al., 2014), followed by additional ephemeral glaciations and warming and cooling cycles leading into continental-scale Antarctic glaciation in the early Oligocene (e.g., Miller et al., 1991; Zachos et al., 1992; Thomas, 1992; Cramer et al., 2011).

Deep-water circulation patterns during this transition changed as key tectonic gateway configurations changed (Exon et al., 2004; Stickley et al., 2004; Scher and Martin, 2006; Livermore et al., 2007; Cramer et al., 2009, 2011; Katz et al., 2011; Borrelli et al., 2014; Borrelli and Katz, 2015), including the progressive openings of the Drake (Cox, 1989; England, 1992; Toggweiler and Bjornsson, 2000; Livermore et al., 2004, 2007; Siijp and England, 2004; Eagles et al., 2006) and Tasman gateways further open in the latest Eocene–early Oligocene (ca. 50 Ma; Scher et al., 2007; Siijp and England, 2004; Eagles et al., 2006) and Tasman gateways opened. Initial opening of the Drake Passage began in the middle Eocene (ca. 50 Ma; Scher et al., 2007). The earliest documented ephemeral Antarctic glaciation occurred in the late-middle Eocene (ca. 37.3 Ma; Scher et al., 2014), followed by additional ephemeral glaciations and warming and cooling cycles leading into continental-scale Antarctic glaciation in the early Oligocene (e.g., Miller et al., 1991; Zachos et al., 1992; Thomas, 1992; Cramer et al., 2011).
ACC in response to the gradual opening of the Drake and Tasman Passages (e.g., Livermore et al., 2007; Cramer et al., 2008; Sijp et al., 2011; Bijl et al., 2013); this is consistent with models that show the formation of cold SCW with the progressive deepening of the Drake and Tasman Passages (e.g., Toggweiler and Samuels, 1995; Toggweiler and Bjørnsson, 2000; Nong et al., 2000; Sijp and England, 2004, 2005). The opening of the Drake Passage and developing proto-ACC (defined as the precursor to the modern ACC; the proto-ACC was shallower than the modern current because of the more constricted Drake and Tasman gateways; e.g., Borrelli et al., 2014) also has been linked to the production of Northern Component Water (NCW; deep-water mass originating in the northern North Atlantic; e.g., Scher and Martin, 2008; Katz et al., 2011; Borrelli et al., 2014) that originated in the late-middle Eocene from the Labrador Sea (Borrelli et al., 2014). In addition, the Tethys closed to the North Atlantic during the middle Eocene, which prohibited export of low-latitude warm saline deep water to the Atlantic Ocean from the western Tethys (Oberhänsli, 1992).

In the marine record, δ18O values of foraminifera are used to interpret paleo-temperature and global ice volume (e.g., Miller and Fairbanks, 1987; Cramer et al., 2009), while δ13C values are indicative of paleoproductivity, water-mass aging, and the oceanic carbon cycle (e.g., Kroopnick, 1985; Lynch-Stieglitz et al., 1995). The analysis of benthic foraminifera that record isotopic changes provides a snapshot of global climate and deep-water circulation from the time they formed their tests. Intersite comparisons between benthic foraminiferal δ18O and δ13C records are used to interpret the source region of deep-water production and the flow path of deep-water masses (e.g., Curry and Lohmann, 1982; Oppo and Fairbanks, 1987; Miller, 1992; Wright et al., 1992; Zachos et al., 2001; Cramer et al., 2009).

The global changes in deep-water circulation at million-year time scales associated with the shift from virtually ice-free conditions during the early-middle Eocene to the development of continental-scale Antarctic glaciation in the early Oligocene has not been documented fully in the South Atlantic. Here we investigate changes in deep-water circulation in the southeastern Atlantic during this transition by comparing new benthic foraminiferal stable oxygen and carbon isotopes (δ18O and δ13C) from Ocean Drilling Program (ODP) Site 1263 (Walvis Ridge, 2700 m present depth, 2100 m paleodepth) and Deep Sea Drilling Project (DSDP) Site 366 (eastern equatorial Atlantic, 2853 m present depth, 2200–2800 m paleodepth) to published data sets from the Atlantic and Southern Oceans (Table 1; Figs. 1 and 2). We compare the circulation changes to gateway changes in order to examine how the development of the proto-ACC affected deep-water circulation in the southeastern Atlantic, as has been suggested for other regions including the Southern Ocean, northern Atlantic, and northeastern Pacific (e.g., Diester-Haass and Zahn, 1996; Diekmann et al., 2004; Scher and Martin, 2006; Via and Thomas, 2006; Cramer et al., 2009; Katz et al., 2011; Borrelli et al., 2014; Borrelli and Katz, 2015).

### MODERN SOUTH ATLANTIC OCEAN CIRCULATION

The modern ocean is driven by thermohaline circulation (e.g., Gordon, 1986; Broecker, 1991; Jayne and Marotzke, 2001), wind-driven surface mixing (Wyrtki, 1961; Schmitz, 1995; Rahmstorf, 2003), and eddy diffusivity (e.g., Holzer and Primeau, 2006). The Meridional Overturning Circulation (MOC) in the present-day ocean is driven primarily by wind-induced upwelling that is most intense along the ACC and causes the displacement of deep water to the surface, allowing formation of new deep water through downwelling in the North Atlantic and Southern Ocean (e.g., Toggweiler and Samuels, 1998). The South Atlantic plays an important role in the global balance of the modern Atlantic Meridional Overturning Circulation (AMOC) through mixing, advection, and subduction of water masses that enter this basin from all depths and oceans (Garzoli and Matano, 2011). The AMOC is potentially a strong influence on today’s global climate because of its large interbasin transfer of heat (Garzoli and Matano, 2011).

Today, Site 1263 (Walvis Ridge) is influenced by a combination of Circumpolar Deep Water (CDW) mixed with North Atlantic Deep Water from the north and west, and CDW that travels north directly from the ACC (Fig. 1; Friedrichs et al., 1994; Speer et al., 1995; Stramma and England, 1999). The mid-Atlantic, Walvis, and Agulhas Ridges have constrained deep-water circulation in the southeastern Atlantic since ca. 130 Ma (O’Connor and Duncan, 1990; Müller et al., 1993; Stramma and England, 1999; Fig. 1). From north to south, these ridges mark the boundaries for the Angola, Cape, and Agulhas Basins.

#### TABLE 1. PALEODEPTHS OF DEEP SEA DRILLING PROJECT AND OCEAN DRILLING PROGRAM SITES DISCUSSED IN THE TEXT

<table>
<thead>
<tr>
<th>Location</th>
<th>Middle to late Eocene paleodepth (m)</th>
<th>Citation</th>
</tr>
</thead>
<tbody>
<tr>
<td>ODP Site 1263 Walvis Ridge</td>
<td>-2100</td>
<td>Zachos et al. (2004)</td>
</tr>
<tr>
<td>DSDP Site 366 Sierra Leone Rise</td>
<td>-2200–2800</td>
<td>Shipboard Scientific Party (1978)</td>
</tr>
<tr>
<td>ODP Site 689 Maud Rise</td>
<td>-1400</td>
<td>Kennett and Stott (1990)</td>
</tr>
<tr>
<td>ODP Site 699 East Georgia Basin</td>
<td>-3400</td>
<td>Mead et al. (1993)</td>
</tr>
<tr>
<td>ODP Site 1090 Agulhas Ridge</td>
<td>-3200</td>
<td>Pusz et al. (2009, 2011)</td>
</tr>
<tr>
<td>ODP Site 1053 Blake Nose</td>
<td>-1500</td>
<td>Katz et al. (2011); Borrelli et al. (2014)</td>
</tr>
</tbody>
</table>

Note: ODP—Ocean Drilling Program; DSDP—Deep Sea Drilling Project.
Tectonic changes have affected oceanographic circulation at >10^9 yr time scales. The separation of Antarctica from South America and Australia led to the formation of the ACC, which may have caused thermal isolation of the Southern Ocean as early as the latest Eocene by preventing the flow of warm surface currents from the subtropical gyres to Antarctica (Toggweiler and Bjornsson, 2000; Barker, 2001; Exon et al., 2004; Sijp and England, 2004, 2005). Paleooceanographic evidence indicates that surface flow through the Tasman Passage was in place during the middle Eocene and resulted in cooling of the Southern Ocean earlier than previously reported (Exon et al., 2004; Stickley et al., 2004) for the late Eocene (Bijl et al., 2013). Plate tectonic models indicate that the Drake Passage began to open ca. 50 Ma and may have contributed to cooling and growth of ice sheets in Antarctica during the middle Eocene (Livermore et al., 2007; Cramer et al., 2011). Before a deep and continuous ACC was established, flow of the shallower proto-ACC was likely controlled by tectonic changes that either constricted or enabled Drake Passage throughflow (Livermore et al., 2004, 2007). In addition, surface water originating from the Pacific and Tethys may have downwelled and traveled as deeper currents into the North and South Atlantic Basins (Kennett and Stott, 1990; Barrera and Huber, 1993; Mead et al., 1993; Thomas, 2005; Scher and Martin, 2006; Via and Thomas, 2006). The closing of the Tethys may have occurred as early as 35 Ma (Allen and Armstrong, 2008), and would have restricted flow from the Tethys into the North Atlantic and changed the water-mass characteristics in the southeastern Atlantic (Allen and Armstrong, 2008). This restriction of deep-water flow from the Tethys to the North Atlantic may have amplified Tethyan flow eastward into the Indian Ocean, much like the modern-day Red Sea Intermediate Water into the Agulhas Current (e.g., Roman and Lutjeharms, 2009). The closing of the Tethys to the North Atlantic, combined with the development of the proto-ACC, established a tectonic configuration similar to today. The ACC is an important component of modern-day AMOC by controlling wind-driven mixing, a key component of meridional ocean circulation (Kuhlbrodt et al., 2007; Garzoli and Matano, 2011); however, the role of the ACC in setting the strength of the northern cell of the modern MOC is still a matter of debate (Kuhlbrodt et al., 2007). In addition, the Agulhas leakage, which consists of rings broken off from the Agulhas Current at the Agulhas retroflection off the southern tip of the African peninsula, can affect AMOC by altering the flow of warm, salty Indian Ocean water into the South Atlantic (van Leeuwen et al., 2000; Speich et al., 2007; Biastoch et al., 2009; Garzoli and Matano, 2011).
In this paper we explore the evolution of deep-water flow and early AMOC at the Walvis Ridge (eastern South Atlantic) in the late-middle Eocene to the earliest Oligocene (40–33 Ma), during the early stages of the opening of the Drake and Tasman Passages and development of the ACC. We monitor changes in water masses at the Walvis Ridge by comparing benthic foraminiferal δ18O and δ13C from Sites 1263 and 366 to published records from other regions within the Atlantic Ocean (Table 1; Figs. 1 and 2). Together, these new and published records provide the means to reconstruct deep-water circulation in the Atlantic Ocean and provide evidence of meridional overturning circulation in the South Atlantic as it responded to the development of the proto-ACC during the period leading up to the large-scale glaciation of Antarctica.

### METHODS

**ODP Site 1263,** Hole B (Site 1263; 28°30’S, 2°45’E) was drilled at 2171 m water depth on the Walvis Ridge (Figs. 1 and 2). The paleodepth for this location from the middle Eocene to early Oligocene was ~2100 m, with general lithology consisting of nannofossil ooze, chalky nannofossil ooze, foraminifera-bearing nannofossil ooze, and clay-bearing nannofossil ooze (Zachos et al., 2004). DSDP Site 366 (Figs. 1 and 2) is located in the eastern equatorial Atlantic on the Sierra Leone Rise, 4°40.70’N, 19°51.10’W, ~800 km west of Sierra Leone with a present water depth of 2853 m (Shipboard Scientific Party, 1978). The Site 366 paleolatitude during the Eocene was ~2°N, with a paleodepth of 2200–2800 m during the middle Eocene to early Oligocene (Miller et al., 1989). The general lithology of this section at Site 366 consists of nannofossil ooze and/or chalk, pelagic clay, and siliceous limestone (Shipboard Scientific Party, 1978).

#### Sample Preparation

A total of 336 samples from ODP Site 1263B were obtained at 15 cm intervals and 157 samples were obtained from DSDP Site 366. Samples from Site 366 were soaked in a sodium metaphosphate solution for no more than 24 h to prevent initial chemical alteration. N. truempyi specimens from each sample from Sites 1263 and 366 were analyzed in the stable isotope laboratory in the Department of Earth and Planetary Sciences at Rutgers University (New Jersey) (Supplementary Tables 1 and 2; Fig. 3). Foraminifera were reacted in phosphoric acid at 90 °C for 15 min in an automated peripheral attached to a Micromass Optima mass spectrometer.

The equations listed here were used for the interspecies calibrations for *Cibicidoides* spp. with *O. umbonatus* and *N. truempyi* for analyses from Site 1263. Equations to calibrate *N. truempyi* (Sites 1263 and 688) δ18O and δ13C, and *O. umbonatus* δ18O (Site 1263), are from Katz et al. (2003). The equation to calibrate *O. umbonatus* δ13C (Site 1263) is from Shackleton et al. (1984) because it provides the best fit to the *Cibicidoides* spp. δ13C data as compared to the calibration in Katz et al. (2003). These calibrated values are within the range of the *Cibicidoides* values (Fig. 3). Other published data used in this study utilized *Cibicidoides*, and needed no calibration.

1. **O. umbonatus:**
   \[ \delta^{18}O: y = x + 0.28 \]
   such that *Cibicidoides* spp. δ18O = (O. umbonatus δ18O) – 0.28
   \[ \delta^{13}C: y = x – 1.00 \]
   such that *Cibicidoides* spp. δ13C = (O. umbonatus δ13C) + 1.00

2. **N. truempyi:**
   \[ \delta^{18}O: y = 0.89x – 0.10 \]
   such that *Cibicidoides* spp. δ18O = [(N. truempyi δ18O) + 0.10]/0.89
   \[ \delta^{13}C: y = x – 0.34 \]
   such that *Cibicidoides* spp. δ13C = (N. truempyi δ13C) + 0.34

At Site 366, we primarily used *C. eocaenus* because of its high abundance and excellent preservation. Well-preserved *C. praemundulius, C. bradyi, C. grimsdalei,* and *C. mexicanus* were substituted as needed.

Stable isotope analyses for Site 366 were conducted using a GV Instruments IsoPrime dual inlet mass spectrometer at the University of South Carolina. Stable isotope values are reported versus Vienna Peedee belemnite by analyzing NBS-19 and an internal lab standard during each automated run. We use the published values for NBS-19 of –2.20 and 1.95‰ for δ18O and δ13C, respectively (Coplen, 1995). The 1σ precision of the standards analyzed for Site 366 (~395–475 m below seafloor, mbsf) is shallower than the ~500–550 mbsf at which burial diagenesis may begin to affect δ18O values, with the potential to overprint the primary signal with anomalously low values (Miller and Fairbanks, 1987). The absence of anomalous δ18O values in the lowest part of our record is consistent with preservation of the primary δ18O signal (Fig. 3B).

1. **C. eocaenus**
   \[ \delta^{18}O: y = 0.91x + 0.19 \]
   \[ \delta^{13}C: y = 0.89x + 0.01 \]

1. **C. grimsdalei**
   \[ \delta^{18}O: y = 0.89x + 0.11 \]
   \[ \delta^{13}C: y = 0.91x + 0.03 \]

1. **C. bradyi**
   \[ \delta^{18}O: y = 0.89x + 0.10 \]
   \[ \delta^{13}C: y = 0.91x + 0.06 \]

1. **C. mexicanus**
   \[ \delta^{18}O: y = 0.89x + 0.12 \]
   \[ \delta^{13}C: y = 0.91x + 0.01 \]

1. **C. eocaenus**
   \[ \delta^{18}O: y = 0.89x + 0.11 \]
   \[ \delta^{13}C: y = 0.91x + 0.03 \]

1. **C. grimsdalei**
   \[ \delta^{18}O: y = 0.89x + 0.10 \]
   \[ \delta^{13}C: y = 0.91x + 0.06 \]

1. **C. bradyi**
   \[ \delta^{18}O: y = 0.89x + 0.12 \]
   \[ \delta^{13}C: y = 0.91x + 0.01 \]

1. **C. mexicanus**
   \[ \delta^{18}O: y = 0.89x + 0.11 \]
   \[ \delta^{13}C: y = 0.91x + 0.03 \]
et al. (1989) data and our new analysis of 0.174‰ \( \delta^{18}O \) and 0.138‰ \( \delta^{13}C \). Our higher resolution data set adds detailed structure to the lower resolution Miller et al. (1989) record, and shows that it was subject to signal aliasing as a result of low resolution.

**Age Models**

The ages for datums in all age models were updated to the 2004 geological time scale (2004 GTS; Gradstein et al., 2005; Tables 2A–2F). We use this time scale to allow comparisons to published data from the compilation of Cramer et al. (2009). Ages for the planktonic foraminiferal datums are from Wade et al. (2011) and the ages of late Eocene to early Oligocene calcareous nannoplankton are from Blaj (2009).

**Site 1263**

Most of the sediment in Hole 1263B from the middle Eocene to lower Oligocene is dominated by nanofossil oozes, which did not preserve a strong magnetic signal in much of this section (Zachos et al., 2004). Nonetheless, the base of magnetochron C13n was identified in Hole 1263B (Bowles, 2006) and is used in our age model, supported by the oxygen isotope correlation with Site 889. The highest occurrence (HO) of the nanoplankton *Ericsonia formosa* was used as the uppermost datum in the core (Fig. 4A) because it is well constrained (Blaj, 2009).

We identified the last occurrence of *Hantkenina* in our samples from Site 1263 at 98.89 m composite depth (mcd), updating the depth of 104.05 mcd provided by the ODP Site 1263 initial report, which was based on lower resolution samples (Zachos et al., 2004). The age of this datum was updated from 33.7 to 33.9 Ma from the 1995 geologic time scale (Cande and Kent, 1995) to the 2004 GTS (Gradstein et al., 2005; Wade et al., 2011); the use of this datum in the age model is supported by the \( \delta^{18}O \) correlation between Sites 1263 and 889 (Fig. 5). The planktonic foraminifer *Turborotalia cerroazulensis* was not used in our age model because specimens were rare, with poorly defined first and last occurrences at Site 1263 (Zachos et al., 2004); in addition, its range was sporadic and not ubiquitous among various global locations (Berggren and Pearson, 2005). The unreliability of *T. cerroazulensis* at Site 1263 is supported by the fact that if it is included in the age model, it results in an unreasonably old age for Oligocene oxygen isotope event 1 (Oi-1; e.g., Miller et al., 1991,
### TABLE 2A. DATUMS USED TO DEVELOP THE AGE MODELS FOR OCEAN DRILLING PROGRAM SITE 1263

<table>
<thead>
<tr>
<th>Depth (mcd)</th>
<th>Age (Ma)</th>
<th>Datum Used in age model</th>
<th>Hole used for datum</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>83.62</td>
<td>33.266</td>
<td>T C13n</td>
<td>1263A</td>
<td>Bowles (2006)</td>
</tr>
<tr>
<td>86.00</td>
<td>32.919</td>
<td>HO Ericsonia formosa</td>
<td>x</td>
<td>1263B</td>
</tr>
<tr>
<td>91.74</td>
<td>33.738</td>
<td>B C13n</td>
<td>x</td>
<td>Blaj (2009)</td>
</tr>
<tr>
<td>98.05</td>
<td>34.000</td>
<td>HO Turborotalia cermaena</td>
<td>x</td>
<td>1263B</td>
</tr>
<tr>
<td>98.89</td>
<td>33.900</td>
<td>HO Hankenina spp.</td>
<td>x</td>
<td>Bowles (2006)</td>
</tr>
<tr>
<td>104.30</td>
<td>34.435</td>
<td>HO Discoster saipanensis</td>
<td></td>
<td>1263B</td>
</tr>
<tr>
<td>104.05</td>
<td>34.500</td>
<td>HO Acarinina primitiva</td>
<td>x</td>
<td>Zachos et al. (2004)</td>
</tr>
<tr>
<td>138.47</td>
<td>39.370</td>
<td>MECO</td>
<td>1263B</td>
<td>This study</td>
</tr>
<tr>
<td>148.01</td>
<td>40.440</td>
<td>T C19n</td>
<td>x</td>
<td>Zachos et al. (2004)</td>
</tr>
</tbody>
</table>

*Note: Core depths are in meters composite depth (mcd). HO—highest occurrence; LO—lowest occurrence; T—top of magnetochron; B—base. MECO—Middle Eocene Climatic Optimum.*

### TABLE 2B. DATUMS USED TO DEVELOP THE AGE MODELS FOR DEEP SEA DRILLING PROJECT SITE 366

<table>
<thead>
<tr>
<th>Depth (mbsf)</th>
<th>Age (Ma)</th>
<th>Datum</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>382.34</td>
<td>32.00</td>
<td>HO Pseudohastigerina spp.</td>
<td>Miller et al. (1989)</td>
</tr>
<tr>
<td>406.33</td>
<td>33.65</td>
<td>δ13C maximum</td>
<td>This study</td>
</tr>
<tr>
<td>462.87</td>
<td>38.00</td>
<td>E14-E15 boundary</td>
<td>Krasheninnikov and Pflaumann (1978)</td>
</tr>
</tbody>
</table>

*Note: mbsf—meters below seafloor. HO—highest occurrence. E is planktonic foraminiferal biozone.*

### TABLE 2C. DATUMS USED TO DEVELOP THE AGE MODELS FOR OCEAN DRILLING PROGRAM SITE 689

<table>
<thead>
<tr>
<th>Depth (mbsf)</th>
<th>Age (Ma)</th>
<th>Datum</th>
<th>(magnetochron)</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>90.20</td>
<td>28.186</td>
<td>T C10n.1n</td>
<td></td>
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<tr>
<td>92.10</td>
<td>28.715</td>
<td>B C10n.2n</td>
<td></td>
<td></td>
</tr>
<tr>
<td>104.50</td>
<td>30.627</td>
<td>T C12n</td>
<td></td>
<td></td>
</tr>
<tr>
<td>116.70</td>
<td>33.266</td>
<td>T C13n</td>
<td></td>
<td></td>
</tr>
<tr>
<td>120.20</td>
<td>33.738</td>
<td>B C13n</td>
<td></td>
<td></td>
</tr>
<tr>
<td>124.10</td>
<td>35.266</td>
<td>B C16n.2n</td>
<td></td>
<td></td>
</tr>
<tr>
<td>144.40</td>
<td>37.235</td>
<td>B C17n.1n</td>
<td></td>
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<tr>
<td>197.50</td>
<td>49.500</td>
<td>hiatus</td>
<td></td>
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</tr>
</tbody>
</table>

*Note: mbsf—meters below seafloor. Data from Bohaty and Zachos (2003); Diester-Haass and Zahn (1996), Ages updated to the 2004 geological time scale (Gradstein et al., 2005). T—top; B—base.*

### TABLE 2D. DATUMS USED TO DEVELOP THE AGE MODELS FOR OCEAN DRILLING PROGRAM SITE 699

<table>
<thead>
<tr>
<th>Depth (mbsf)</th>
<th>Age (Ma)</th>
<th>Datum</th>
<th>(magnetochron)</th>
<th>Datum</th>
</tr>
</thead>
<tbody>
<tr>
<td>253.64</td>
<td>31.26</td>
<td>B hiatus</td>
<td></td>
<td></td>
</tr>
<tr>
<td>284.55</td>
<td>33.738</td>
<td>B C13n</td>
<td></td>
<td></td>
</tr>
<tr>
<td>329.20</td>
<td>36.000</td>
<td>LO Isthmolithus recurvis</td>
<td></td>
<td></td>
</tr>
<tr>
<td>355.20</td>
<td>39.400</td>
<td>MECO</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

*Note: mbsf—meters below seafloor. MECO—Middle Eocene Climatic Optimum. LO—lowest occurrence; C—magnetochron; B—base. Data from Mead et al. (1993). Ages updated to the 2004 geological time scale (Gradstein et al., 2005).*

### TABLE 2E. DATUMS USED TO DEVELOP THE AGE MODELS FOR OCEAN DRILLING PROGRAM SITE 1053

<table>
<thead>
<tr>
<th>Depth (mbsf)</th>
<th>Age (Ma)</th>
<th>Datum</th>
<th>(magnetochron)</th>
<th>Datum</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.00</td>
<td>33.900</td>
<td>HO Hankenina</td>
<td></td>
<td></td>
</tr>
<tr>
<td>19.52</td>
<td>35.040</td>
<td>B C15n</td>
<td></td>
<td></td>
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<tr>
<td>80.73</td>
<td>36.000</td>
<td>LO Isthmolithus recurvis</td>
<td></td>
<td></td>
</tr>
<tr>
<td>144.83</td>
<td>37.240</td>
<td>B C17n.1n</td>
<td></td>
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<td>160.41</td>
<td>37.550</td>
<td>B C17n.2n</td>
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<tr>
<td>172.57</td>
<td>37.700</td>
<td>HO Morozovella spinulosa</td>
<td></td>
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<tr>
<td>182.16</td>
<td>38.000</td>
<td>LO Globigerinatha semiinvoluta</td>
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*Note: mbsf—meters below seafloor. HO—highest occurrence; LO—lowest occurrence. C—magnetochron; T—top; B—base. Data from Borrelli et al. (2014). Ages updated to the 2004 geological time scale (Gradstein et al., 2005).*

### TABLE 2F. DATUMS USED TO DEVELOP THE AGE MODELS FOR OCEAN DRILLING PROGRAM SITE 1090

<table>
<thead>
<tr>
<th>Depth (mbsf)</th>
<th>Age (Ma)</th>
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<tr>
<td>118.37</td>
<td>29.740</td>
<td>B C11n</td>
<td></td>
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<tr>
<td>119.36</td>
<td>30.627</td>
<td>T C12n</td>
<td></td>
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<tr>
<td>124.89</td>
<td>31.116</td>
<td>B C12n</td>
<td></td>
<td></td>
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<tr>
<td>139.25</td>
<td>33.266</td>
<td>T C13n.1n</td>
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<td>146.64</td>
<td>33.738</td>
<td>B C13n.2n</td>
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<td>155.70</td>
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<td>158.10</td>
<td>35.043</td>
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<td>161.60</td>
<td>35.404</td>
<td>T C16n.1n</td>
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<td>162.90</td>
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<td>167.79</td>
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<tr>
<td>172.85</td>
<td>36.600</td>
<td>B C16n.3n</td>
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</tr>
</tbody>
</table>

*Note: Ages updated to the 2004 geological time scale (Gradstein et al., 2005). mbsf—meters below seafloor. T—top; B—base. Data from Pusz et al. (2009, 2011).*
Figure 4. (A) Ocean Drilling Program Site 1263 age-depth plot (mcd—meters composite depth) (see Table 2A). Datums used in age model are shown in red and are connected. Unused datums are in black. HO—highest occurrence, LO—lowest occurrence, T—top, B—base. Sedimentation rates (m/m.y.) were calculated by linear interpolation between datums, and are shown with associated line segment. C—magnetochron; MECO—middle Eocene climatic optimum; T— Turborotalia; G.—Globigerinatheka; E.—Ericsonia; D.—Discoaster; A.—Acarinina; C.—Chiasmolithus; I—Ishmilithus. (B) Deep Sea Drilling Project Site 366 age-depth plot (mbsf—meters below seafloor) (see Table 2B). Core recovery is shown to highlight the reason for data gaps. Sedimentation rates (m/m.y.) were calculated by linear interpolation between datums, and are shown with associated line segment. E15(B)—base of planktonic foraminiferal biozone.

The age model for DSDP Site 366 (Fig. 4B; Table 1) is based on the highest occurrence of *Pseudothastigerina* spp. (Krasheninnikov and Pflaumann, 1978; Miller et al., 1989), adjusted to the 2004 GTS, a tie point with the average age of the peaks in δ¹⁸O records from Sites 1263 and 689 at the Eocene-Oligocene transition, and the base of the E15 planktonic foraminiferal biozone (Wade et al., 2011). Sparse age control datums may result in uncertainty, especially in the 420–450 m below seafloor (mbsf) interval. Coring gaps may add additional uncertainty: protocol is to move all recovered sediment to the top of the drilled interval, even though it is possible that the recovered sediment should have been placed lower in the drilled interval. This is an issue especially for cores 11 (423.0–432.5 mbsf, recovery 1.65 m), 12 (432.5–442.0 mbsf, recovery 5.8 m), and 13 (442.0–452.5 mbsf, recovery 3.4 m; Fig. 4B). To explore the impact of following the convention of moving all recovered sediment to the top of each drilled interval, we moved each sediment core to the bottom of each drilled interval and ran an alternate age model. This also resulted in moving the datums used in the age model, and we emphasize that because each core was moved a different amount (because each core gap is different), the resulting changes are nonlinear, with some data from some cores becoming older than the original, while others become younger. A comparison of both versions of the Site 366 isotope record with the other isotope records used in this study shows that the timing of some of the Site 366 isotope changes shift, but the overall interpretations of the data as discussed in the following remain unchanged (Supplementary Fig. 1’).

**Published Isotopic Records**

The age models from Sites 1090 (Pusz et al., 2009, 2011), 1053 (Borrelli et al., 2014), and 689 (Mead et al., 1993) were updated to the 2004 GTS from the middle Eocene to the early Oligocene in order to compare results to Sites 1263 (this study) and 689 (Diester-Haass and Zahn, 1996; Bohaty et al., 2012; the age model was updated to the 2004 GTS used by Cramer et al., 2009).
The δ¹⁸O minima from Sites 689 and 1263 record the MECO (Figs. 5 and 6). Recalibration of the Site 689 age model from the 1995 GTS to the 2004 GTS shows that MECO occurred at 39.4 Ma, which is supported by well-constrained biostratigraphic datums (Cande and Kent, 1995; Gradstein et al., 2005; Cramer et al., 2009). This age for the MECO was used as a tie point with the δ¹⁸O minima from Site 1263. Constant sedimentation rates were assumed between datums, and linear interpolation was used to determine sample ages (Fig. 4).

The δ¹³C and δ¹⁸O records from Ocean Drilling Program (ODP) Walvis Ridge Site 1263 (this study; N. truempyi and O. umbonatus values calibrated to Cibicidoides spp. samples; see text and Fig. 2), Sierra Leone Rise Deep Sea Drilling Project Site 366 (this study), Southern Ocean ODP Site 689 (Kennett and Stott, 1990; Diester-Haass and Zahn, 1996), South Atlantic ODP Sites 699 (Mead et al., 1993) and 1090 (Pusz et al., 2009, 2011), and western North Atlantic Site 1053 (Katz et al., 2011; Borrelli et al., 2014). The Gradstein et al. (2005) geologic time scale is used for the age model. Dashed lines indicate key time periods (see Results and Discussion in text). Stable isotope data are compared to tectonic gateway event reconstructions (Livermore et al., 2007; Stickley et al., 2004). MECO—Middle Eocene Climatic Optimum; Oi-1—Oligocene oxygen isotope event 1.

### RESULTS

**Site 1263 Benthic Foraminiferal δ¹⁸O**

Site 1263 δ¹⁸O showed variability similar to that of Site 689 (Diester-Haass and Zahn, 1996; Bohaty et al., 2012) from the middle Eocene to the early Oligocene, with small offsets at times (Fig. 5). A >1.0‰ increase in δ¹⁸O over ~400 k.y. followed the MECO event at 39.4 Ma at Site 1263. The δ¹⁸O values at Sites 1053
and 699 diverged from Site 1263 beginning ca. 38.2 Ma. The δ¹⁸O values at Sites 1263 and 366 from 38 to 37.6 Ma were between the higher δ¹⁸O values at Site 689 and the lower δ¹⁸O values at Sites 1053 and 699. The δ¹³C records from Sites 689 and 1263 from ca. 37.6–35.8 Ma were within 0.2‰ of each other (Fig. 5).

From ca. 35.8 to 34.2 Ma, the δ¹⁸O at Site 1263 was 0.1‰–0.3‰ higher than at Site 689. The δ¹³C values for this period of time were determined directly from Cibicidoides spp. and from δ¹³C adjusted from O. umbonatus specimens.

The δ¹⁸O values at Site 366 were high and similar to Site 1263 from 36.7 to 35.8 Ma. Site 366 values decreased to a range of ~1‰–0.5‰ from ca. 35 to 34.3 Ma, similar to the δ¹³C records at Sites 1053, 699, and 1090. These records were 0.3‰–0.8‰ lower than Site 1263 until ca. 34 Ma. Just before Oi-1, all sites displayed a rapid and brief decrease in δ¹⁸O values by 0.5‰–1.5‰ ca. 33.9–33.8 Ma. All of the δ¹³C records discussed here increased beginning ca. 34 Ma, which was the start of Oi-1, but Sites 366, 1090, and 699 were ~0.2‰–0.5‰ lower than Site 1263. The δ¹³C values at Sites 1263, 366, 689, and 1090 peaked ca. 33.7 Ma. Site 699 δ¹³C remained ~0.5‰ lower and Site 366 δ¹³C remained ~1‰ lower than at Site 1263 in the earliest Oligocene.

### Site 1263 Benthic Foraminiferal δ¹³C

The record of benthic foraminiferal δ¹³C at Site 1263 was most similar to the records from Sites 1053 and 699, while the δ¹³C records from Sites 689, 1090, and 366 were less similar to Site 1263 during the time span covered here (Fig. 5). The Site 1263 and Site 699 δ¹³C values were generally 0.5‰ to >1‰ higher than at Site 689 from 39.6 to 36.2 Ma (Fig. 5); ca. 36.2 Ma, Site 689 values became similar to those of Sites 1263 and 1053. Sites 1053 and 1263 had similar δ¹³C records from ca. 375 to ca. 35.8 Ma, but with greater short-term

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**Figure 6. Deep-water evolution based on benthic foraminiferal stable isotope data.** The δ¹⁸O and δ¹³C (‰) changes at Ocean Drilling Program Sites 1263, 689, 1053, 699, and 1090 and Deep Sea Drilling Project Site 366 from 39 Ma to 34 Ma in paleodepth versus paleolatitude at 1 m.y. intervals are shown. Solid lines designate different water masses; dashed lines designate possible mixing of water masses. See Figure 4 caption for references. SCW—Southern Component Water, NCW—Northern Component Water, WSDW—warm saline deep water.
variability at ODP Site 1053. The $\delta^{13}C$ at Site 689 increased by $>$0.5‰ from 36.8 to 36.2 Ma and was similar to Site 1263 during this time and until ca. 35.2 Ma. The $\delta^{13}C$ at Site 1090 was significantly lower than the other sites from ca. 36.2 to 34 Ma. Site 1263 $\delta^{13}C$ values became $-0.2%_{\circ}$–$0.5%_{\circ}$ higher than those of Sites 1053 and 689 from ca. 35 to 34.2 Ma, with convergence of these records ca. 34.2 Ma. The variability at Sites 689 and 1263 was similar from 34 to 33 Ma, although Site 1263 $\delta^{13}C$ was $>0.5%_{\circ}$ higher than at Site 689. The $\delta^{13}C$ at Sites 689 and 1263 increased and peaked between 33.7 and 33.6 Ma. The peak values at Sites 1263 and 689 were $-2.3%_{\circ}$ and $-1.6%_{\circ}$, respectively.

Site 366 Benthic Foraminiferal $\delta^{18}O$

The benthic foraminiferal $\delta^{18}O$ at Site 366 in the oldest part of this record (ca. 39–38.8 Ma) was similar to Sites 1263 and 689, $-1%_{\circ}$ (Fig. 5). Values ranged from $-0.5%_{\circ}$ to $1%_{\circ}$ from ca. 38.1 to 37.2 Ma. These values were $-0.2%_{\circ}$–$0.5%_{\circ}$ lower than at Sites 689 and 1263 and were similar to $\delta^{18}O$ at Sites 1053 and 699.

The $\delta^{18}O$ at Site 366 was $-1.3%_{\circ}$–$1.5%_{\circ}$ from ca. 36.8 to 35.6 Ma and was similar to values at Site 1263. The $\delta^{18}O$ at both of these sites was $-0.1%_{\circ}$–$0.2%_{\circ}$ higher than values at Site 689 for much of this time. Due to coring gaps and the uncertainty of the Site 366 age model (see preceding), we are unable to precisely locate the timing of the $\delta^{18}O$ increase at Site 366 for this million-year period. The $\delta^{18}O$ values at Site 366 were low and more similar to those of Sites 1053, 1090, and 699 during the latter part of the record, from ca. 35 to 34 Ma.

The high variability in $\delta^{18}O$ and $\delta^{13}C$ values in portions of the records shown in this study (both new and published data) has been documented at middle Eocene sites in different basins (e.g., northern Pacific ODP Site 884, Pak and Miller, 1995; Borrelli and Katz, 2015; subtropical Pacific ODP Site 1209, Dawber and Tripati, 2011; equatorial Pacific ODP Sites 1218 and 1219, Tripati et al., 2005; tropical western Atlantic ODP Sites 1258 and 1260, Sexton et al., 2006; northwestern Atlantic Site 1053, Katz et al., 2011; Borrelli et al., 2014; Atlantic and Indian sector of the Southern Ocean, ODP Sites 689, 738, and 748, Bohaty and Zachos, 2003).

Site 366 Benthic Foraminiferal $\delta^{13}C$

The earliest part of the Site 366 $\delta^{13}C$ record ca. 39–38.8 Ma had values that were $-0.5%_{\circ}$–$1%_{\circ}$ lower than at Site 689 and $-1%_{\circ}$–$1.5%_{\circ}$ lower than at Site 1263 (Fig. 5). Site 366 $\delta^{13}C$ values are the lowest of any of the sites from 36.2 to 35.7 Ma. The $\delta^{13}C$ remained low throughout the majority of this record ($0.5%_{\circ}$–$1%_{\circ}$), and the trend did not resemble other sites until the latest Eocene to earliest Oligocene, when the $\delta^{13}C$ values increased at all sites shown here by $-1%_{\circ}$.

### DISCUSSION

We compare the Walvis Ridge Site 1263 and Sierra Leone Rise Site 366 benthic foraminiferal $\delta^{18}O$ and $\delta^{13}C$ (this study) to coeval published records from the Atlantic and Southern Oceans (Fig. 5). In the following sections, we explore the evolution of South Atlantic deep-water circulation during the late-middle Eocene to the earliest Oligocene within the context of these isotopic relationships. For purposes of discussion, we separate time intervals into segments that reflect major variations in the deep-water characteristics at Site 1263: ca. 40–38.2 Ma, ca. 38.2–37.6 Ma, ca. 37.6–36.6 Ma, and ca. 36.6 to the earliest Oligocene (Figs. 5 and 6). A summary of the interpretations from our results is provided in Table 3.

<table>
<thead>
<tr>
<th>Age interval (Ma)</th>
<th>Results</th>
<th>Interpretations</th>
</tr>
</thead>
<tbody>
<tr>
<td>36.60–33.00</td>
<td>Highest $\delta^{18}O$ and $\delta^{13}C$ values at ODP Sites 1263, 689, and DSDP Site 366. Lowest $\delta^{18}O$ values at Sites 1053, 699, and 1090. Site 366 $\delta^{18}O$ values decreased, becoming similar to Site 1053 by 35.6–35 Ma.</td>
<td>SCW at Sites 1263 and 366 at 36.6 Ma, replaced by NCW at Site 366 by 35 Ma, indicating greater NCW production and/or a decrease in SCW production around this time.</td>
</tr>
<tr>
<td>37.60–36.60</td>
<td>Highest $\delta^{18}O$ values at Site 1263, relatively high values at Site 689, and relatively low values at Sites 366, 1053, and 699. $\delta^{18}O$ values at Site 366 increased to values similar to Site 1263 at the end of this interval. Relatively low $\delta^{13}C$ values at Sites 689 and 366.</td>
<td>Greater SCW production with stratification of colder water at greater depths and in the equatorial Atlantic. WSDW still present at greater depths in the South Atlantic Sites 699 and 1090 (mixed with PW at Site 1090). Warm NCW at Site 1053. Higher productivity in the flow paths to Sites 689 and 366 than at other sites listed.</td>
</tr>
<tr>
<td>38.20–37.60</td>
<td>$\delta^{18}O$ values at Sites 699, 366, and 1053 were lower than at Site 689. $\delta^{18}O$ and $\delta^{13}C$ values at Site 1263 were intermediate between these sites.</td>
<td>Possible mixing of water masses in the southeastern Atlantic. Warm water masses present at Sites 1053 (NCW), 699 (WSDW), and 366 (NCW or WSDW).</td>
</tr>
<tr>
<td>40.00–38.20</td>
<td>$\delta^{18}O$ values at Sites 1263, 689, and 699 were similar; $\delta^{13}C$ values were lower at Site 689 than at other sites.</td>
<td>Well-mixed ocean with higher local productivity affecting Site 689 $\delta^{13}C$.</td>
</tr>
</tbody>
</table>

Note: DSDP—Deep Sea Drilling Project; all other sites are Ocean Drilling Program (ODP); SCW—Southern Component Water; NCW—Northern Component Water; WSDW—warm saline deep water; PW—Pacific water.
MECO to 38.2 Ma: Onset of Deep-Water Differentiation

From the beginning of MECO ca. 39.6 Ma until ca. 38.2 Ma, the Walvis Ridge Site 1263 and Sierra Leone Rise Site 366 δ¹⁸O records were very similar to the Southern Ocean Site 689 and East Georgia Basin Site 699, including the MECO δ¹⁸O minimum of ~0.1‰ ca. 39.4 Ma (Figs. 5 and 6). This indicates a uniform deep-water mass bathing several regions of the Atlantic at several paleodepths, as also shown by neodymium isotope data ($e_{Nd}$) from the Walvis Ridge (Sites 1262 and 1263; Via and Thomas, 2006) (Supplementary Fig. 2). In contrast, $e_{Nd}$ data from the Agulhas Ridge (Site 1090; Scher and Martin, 2006) (Supplementary Fig. 2) show the influence of Pacific seawater at this location starting ca. 41.3 Ma (data not included in Supplementary Fig. 2). Without benthic foraminiferal stable isotope data from Site 1090, we can only note that the Atlantic-Pacific δ¹⁸O offset was minimal at this time, consistent with fairly uniform deep-water temperatures (Cramer et al., 2009).

The δ¹⁸O at Sites 1263, 689, and 699 increased after MECO, indicating uniform deep-water cooling; δ¹⁸O at Site 1263 was ~0.5‰–1.0‰ higher than at Site 689 at that time (Figs. 5 and 6A), consistent with paleoproductivity proxies that indicate high productivity at Site 689 (Diester-Haass and Zahn, 1996) and with the location of Site 1263 beneath the South Atlantic gyre at that time. The δ¹⁸O at Sites 1263, 689, and 699 is consistent with (1) paleoproductivity proxies that indicate high productivity at Site 689 (Diester-Haass and Zahn, 1996) and with the location of Site 1263 beneath the South Atlantic gyre at that time.

38.2–37.6 Ma: Deep-Water Mixing in the Southeastern Atlantic

The highest δ¹⁸O was recorded at Maud Rise Site 689 (Kennett and Stott, 1990) and likely resulted from Southern Ocean cooling in response to the onset of thermal isolation with the development of the proto-ACC, which accompanied gateway openings (e.g., Barker, 2001; Exon et al., 2004; Scher and Martin, 2006; Cramer et al., 2009; Katz et al., 2011; Bijl et al., 2013; Borrelli et al., 2014; see Introduction herein for details). Site 689 also had lower δ¹³C, which is consistent with (1) paleoproductivity proxies that indicate high productivity at Site 689 (Diester-Haass and Zahn, 1996), and (2) the location of Site 1263 beneath the South Atlantic gyre at that time. From 38.2 to 37.8 Ma, the δ¹⁸O and δ¹³C values at Walvis Ridge Site 1263 were slightly lower than at Site 689, and higher than at Blake Nose Site 1053 and East Georgia Basin Site 699 (Figs. 5 and 6B). This may indicate that Site 1263 was bathed with (1) SCW that was slightly warmer than recorded at Site 689, or (2) SCW mixed with a warmer deep-water mass, possibly from the Pacific (Scher and Martin, 2006), North Atlantic, or Tethys Sea (Figs. 5 and 6B).

Site 1263 could have been bathed by SCW mixed with Pacific seawater. Site 1263 δ¹⁸O values from 38.2 to 37.6 Ma are similar to Site 1218 δ¹⁸O values (Lear et al., 2004) and $e_{Nd}$ data are similar to those from Site 1090 (Scher and Martin, 2006), consistent with an influx of radiogenic waters from the Pacific into deep-water formation regions of the Southern Ocean as a consequence of the opening of the Drake Passage (Scher and Martin, 2006) (Supplementary Fig. 2). Site 689 recorded a much weaker response to the influx of Pacific water into the Southern Ocean, possibly indicating the isolation of Pacific-influenced deep water to below the depth of Maud Rise (Site 689; Scher and Martin, 2006) (Supplementary Fig. 2). In this scenario, considering that the Site 1263 paleodepth was deeper than Site 689 and shallower than Site 1090, the water mass at Site 1263 might represent the upper branch of a deep-water mass composed of SCW and Pacific seawater. In contrast, Via and Thomas (2006) suggested that the Walvis Ridge was bathed by a single Southern Ocean intermediate and/or deep-water mass until the early Oligocene; however, the low-resolution $e_{Nd}$ records from Sites 1262 and 1263 for this time (Supplementary Fig. 2) do not allow us to draw a firm conclusion about the possibility of Pacific influence at Site 1263.

NCW began to form in the Labrador Sea during the late-middle Eocene (Borrelli et al., 2014). This deep-water mass is characterized at Site 1053 by lower δ¹⁸O and higher δ¹³C than the SCW at Site 689 (Borrelli et al., 2014). The intermediate δ¹⁸O and δ¹³C values at Site 1263 during this time may represent mixing of NCW and SCW. Mixing of these water masses at Site 1263 is consistent with a meridional circulation component and wind-driven advection in the southeastern Atlantic. The restriction of the Tethys seaway to the Atlantic may have driven meridional transport from the western North Atlantic to the eastern South Atlantic, and wind-driven advection may have been possible with the initiation of the proto-ACC as the Drake Passage opened (Allen and Armstrong, 2008).

Alternatively, the warmer deep water at Site 1263 may have resulted from SCW mixing with warm saline deep water (WSDW) that originated from the high-salinity warm waters of the Tethys seaway (Kennett and Stott, 1990; Mead et al., 1993; Wright and Miller, 1993; Scher and Martin, 2004). Tethyan WSDW had low δ¹⁸O and was probably warmer, saltier, and denser than the overlying high δ¹⁸O SCW (Kennett and Stott, 1990; Mead et al., 1993); $e_{Nd}$ values at Site 689 are consistent with Tethyan $e_{Nd}$ values at that time (Scher and Martin, 2004). Temperature and salinity calculations based on δ¹⁸O records establish the density structure that supports the presence of this warmer, more saline deep water underlying colder, fresher deep water in this region in the Eocene; this warm saline deep water may have formed through mixing of cold high-latitude sourced deep water with WSDW sourced from an evaporative semi-enclosed basin (Mead et al., 1993).

The concomitant formation of relatively warm NCW and cooling of SCW beginning ca. 38 Ma is a response to the progressive development of the (proto-) ACC with the opening of the Drake and Tasman Passages; this isolated surface waters to the south of the ACC from the warmer subtropical gyre, cooled the source region of SCW, and resulted in cooling of SCW (e.g., Toggweiler and Bjornsson, 2000; Bijl and England, 2004, 2005; Livermore et al., 2007; Cramer et al., 2009; Katz et al., 2011; Bijl et al., 2013; Borrelli et al., 2014). In addition, small-scale Antarctic glaciation (Browning et al., 1996; Scher et al., 2014) and the development of the proto-ACC (Cramer et al., 2011) may have contributed to increasing δ¹⁸O at Sites 689 and 1263. The isotopic evidence presented here shows that deep water began to cool only at these sites during that time, which indicates that cold deep water was not being produced in large enough quantities to displace WSDW at all depths and locations, while WSDW was still being supplied to the equatorial Atlantic and South Atlantic (Fig. 6B).
Ca. 37.6–36.6 Ma: Initial Cooling of Southeastern Atlantic Deep Water

Site 1263 is characterized by a slightly higher δ¹⁸O compared to all other sites from ca. 37.6 to 36.6 Ma (Figs. 5 and 6C), indicating cold SCW at Site 1263, whereas the deep-water mass in the northwestern Atlantic (Site 1053) likely originated in the Labrador Sea (Borrelli et al., 2014). The warm deep-water mass at Site 699 may have been derived from WSDW (Mead et al., 1993), possibly originating in the Tethys and entering the South Atlantic via the Indian Ocean, similar to the warm, salty Indian Ocean water that enters the South Atlantic today (van Leeuwen et al., 2000; Speich et al., 2007; Biastoch et al., 2009; Garzoli and Matano, 2011). The δ¹⁸O values at the Sierra Leone Rise Site 366 from ca. 37.6 to 37.2 Ma are similar to those at Sites 1053 and 699, and then increase by ca. 36.8 Ma, becoming similar to δ¹⁸O values at Sites 689 and 1263, indicating that SCW dominated the equatorial Atlantic by that time.

The reduced δ¹³C gradient among all sites at 376–37 Ma (Fig. 5) indicates that there was a possible overall increase in productivity within source waters, in the overlying water column, or along the flow paths for Sites 1263, 699, and 1053 at that time. The similar high δ¹³C values at Site 1263, East Georgia Basin Site 699, and Site 1053 from ca. 37 to 36 Ma, coupled with higher δ¹⁸O (−1.0‰–1.7‰) at Site 1263, suggests that the deep-water masses at these sites derived from different surface waters that were exposed to uniform changes in carbon cycling (e.g., Billups et al., 2002).

The isotopic records presented here reflect the development of at least 4 deep-water masses by ca. 376 Ma. This is consistent with hypotheses that the proto-ACC contributed to cold deep-water formation and deep-water circulation changes as early as ca. 38 Ma (Cramer et al., 2009; Borrelli et al., 2014). These deep-water masses include: (1) warm NCW at Site 1053 with low δ¹⁸O and high δ¹³C (Borrelli et al., 2014); (2) cold SCW at Site 689 with high δ¹⁸O and low δ¹³C (Diester-Haass and Zahn, 1996; Bohaty and Zachos, 2003); (3) WSDW at Site 699 with low δ¹⁸O and high δ¹³C (Mead et al., 1993); and (4) cold SCW at Site 1263 with high δ¹⁸O and high δ¹³C (this study), possibly as a consequence of mixing between SCW and NCW, WSDW, or Pacific seawater, as suggested by Site 1090 f° values (Scher and Martin, 2006) (Supplementary Fig. 2).

The cold deep water at Site 1263 may have shared a SCW source with Site 689; however, high productivity at Site 689 may have lowered the δ¹³C values. By ca. 376 Ma, the deep-water mass at the Walvis Ridge (Site 1263) changed from a mixture of SCW with NCW, WSDW, or Pacific seawater to being primarily SCW sourced (Fig. 5). In the eastern equatorial Atlantic (Site 366), mixing of NCW and/or WSDW mixing with SCW eventually shifted to SCW dominance by ca. 37–36.8 Ma (a coring gap prevents precise placement of this transition). This progressive change may have resulted from greater cold water production in the Southern Ocean, consistent with previously published models showing progressively developing proto-ACC in response to gradual opening of the Drake and Tasman Passages, deeper and stronger flow of the proto-ACC, and thermal isolation of the Southern Ocean (e.g., Cramer et al., 2009; Katz et al., 2011; Bijl et al., 2013). These progressive circulation changes are consistent with the development of a weak early AMOC (Figs. 5, 6B, and 6C). This is analogous to the modern AMOC driven by the ACC in the South Atlantic Ocean (Garzoli and Matano, 2011). It is also consistent with modeling studies that find that the Drake Passage opening initiates ACC flow and AMOC-like circulation (Toggweiler and Bjornsson, 2000; Fyke et al., 2015).

36.6–33 Ma: SCW Deep-Water Stratification

From ca. 36.6 to 35.8 Ma, the southeastern Atlantic deep-water mass at Sites 1263 and 366 recorded δ¹⁸O that ranged from ~0.0‰ to 0.2‰ higher than at Maud Rise Site 689, consistent with SCW flowing through the southeastern Atlantic at least as far north as the equatorial region (Figs. 5 and 6D). The 0.2‰ higher δ¹⁸O values at Sites 1263 and 366 likely reflect the slightly colder, denser waters in the deeper portion of SCW. The δ¹⁸O values at Site 1263 continued to be slightly higher than any other published record until ca. 34.2 Ma (Figs. 6E, 6F). The δ¹⁸O values at Site 366 decreased to a range of −1.2‰–0.5‰ between ca. 35.6 and 35 Ma (across a coring gap), which may reflect mixing with the warmer NCW-sourced water recorded at Site 1053. These circulation changes are consistent with the progressive strengthening of the proto-ACC (e.g., Exon et al., 2004; Cramer et al., 2009; Katz et al., 2011). Similar deep-water stratification is present in the modern Atlantic Ocean (Talley et al., 2011).

The δ¹⁸O values at the deeper south Atlantic Sites 1090 and 699 were within a range of −0.4‰–0.7‰ from ca. 36 to 35.4 Ma (Figs. 5 and 6). These low values indicate that there was either a warm deep-water mass restricted to greater depths than at Sites 689 and 1263, or the cold South Atlantic water was confined to west of Site 699 and a warm deep-water mass flowed into the South Atlantic from the Indian Ocean (Figs. 6D, 6E). However, we note that f° values suggest that Pacific seawater was present at Site 1090 (Scher and Martin, 2006) (Supplementary Fig. 2); taken together with the δ¹⁸O record (Pusz et al., 2009, 2011), this may indicate mixing of WSDW and Pacific seawater at that time.

The δ¹³C values at Site 1263 is higher than at Site 689 (Fig. 5) during most of the period investigated, indicating that productivity was lower in the Argentine and Brazil Basins in the South Atlantic. The increase in δ¹³C at Site 689 ca. 36 Ma may indicate a reduction of productivity in this area. This may reflect the intensification of flow through the Drake Passage, enhanced upwelling, and the concentration of nutrients along the northern edge of the proto-ACC, which may have affected productivity at Agulhas Ridge Site 1090, as evidenced by low δ¹³C values, without reaching East Georgia Basin Site 699 (Fig. 6). High primary production is also supported at Site 1090 by barite, carbonate, and phosphorous accumulation (Anderson and Delaney, 2005) and opal accumulation (Diedkamm et al., 2004; Anderson and Delaney, 2005). In addition, low δ¹³C at Site 366 from ca. 36.2–35.8 Ma likely resulted from oxidation of terrestrial-sourced ¹³C-enriched organic carbon (Wagner, 2000). Based on our isotopic comparisons, 3 distinct water masses existed in a stratified ocean by 35 Ma (Fig. 6E): (1) cold SCW at Sites 689 and 1263; (2) warm NCW at Sites 1053 and 366; and (3) WSDW at Sites 689 and 1090, with Pacific seawater influence at Site 1090.
The $\delta^{18}O$ and $\delta^{13}C$ data from Walvis Ridge Site 1263 converge with records from Site 689 in the latest Eocene, ca. 34.2 Ma (Fig. 6f). Limited data show a brief $\delta^{18}O$ decrease of $-0.5$ to $-1.5\%$ ca. 34 Ma at Sites 1263, 366, 689, 699, and 1053; this may indicate that cold deep-water production was interrupted for a very short period, revealing an interruption or slowing of the developing meridional overturning circulation, and that deep-water production and circulation was unstable. However, additional high-resolution data are needed to better constrain the $\delta^{18}O$ decrease and to investigate the possible instability in meridional overturning circulation at this time.

An $-1.5\%$ increase in $\delta^{18}O$ at all sites occurred from the late Eocene to early Oligocene, including the Eocene-Oligocene transitions 1 and 2 and the Oi-1 isotope event, as a consequence of $-2$ to $-4^\circ$C cooling (e.g., Katz et al., 2008; Lear et al., 2008; Pusz et al., 2011) and associated large-scale Antarctic glaciation that caused a 55 to 70 m eustatic fall (e.g., Kennett and Stott, 1990; Diester-Haass and Zahn, 1996; Zachos et al., 2001; Katz et al., 2008; Miller et al., 2009; Cramer et al., 2011). At Oi-1, Site 1090 $\delta^{18}O$ and $\delta^{13}C$ values shifted toward Site 689 and 1263 values, consistent with an intensification of the proto-ACC or a reduction in the Tethys deep-water source.

## SUMMARY AND CONCLUSIONS

Beginning ca. 38.2 Ma, $\delta^{18}O$ at Maud Rise Site 689 diverged from East Georgia Basin Site 699, Sierra Leone Rise 366, and North Atlantic Site 1053; Walvis Ridge Site 1263 $\delta^{18}O$ values were intermediate between these values, indicating possible mixing of water masses. This divergence was likely in response to increased flow through the Drake and Tasman Passages, which initiated thermal isolation of surface waters around Antarctica, the source area of SCW. There was a simultaneous warming in the North Atlantic (Borrelli et al., 2014) and production of WSDW at low latitudes (Kennett and Stott, 1990; Mead et al., 1993; Wright and Miller, 1993; Scher and Martin, 2004).

Our comparisons indicate that by ca. 37.6 Ma there were at least 4 deep-water masses in the Atlantic and Southern Oceans: (1) warm NCW with low $\delta^{18}O$ and high $\delta^{13}C$ (Site 1053, Borrelli et al., 2014); (2) cold SCW with high $\delta^{18}O$ and low $\delta^{13}C$ (Site 689, Diester-Haass and Zahn, 1996; Bohaty and Zachos, 2003); (3) WSDW with low $\delta^{18}O$ (Site 366 low $\delta^{13}C$, this study; Site 699 with high $\delta^{13}C$, Mead et al., 1993); and (4) cold SCW with high $\delta^{18}O$ and high $\delta^{13}C$ (Site 1263, this study).

At 36.6 Ma, the deep-water masses at Sites 1263 and 366 were colder than the water mass at Site 689, likely reflecting the slightly colder, denser lower portion of SCW; this indicates strong SCW production that dominated the southeast and equatorial Atlantic by that time. The strengthening of the proto-ACC is the only known mechanism that may have led to colder water production from the Southern Ocean reaching the depths of Sites 1263 and 366. By ca. 35 Ma, the $\delta^{18}O$ at Site 366 was substantially lower than at Sites 1263 and 889; this may have resulted from increased production of NCW and/or change of deep-water flow patterns into the equatorial Atlantic.

Gateway openings and closings in the Eocene–early Oligocene affected SCW production and ocean circulation in the South Atlantic Ocean. With a closed Drake Passage, the eastern South Atlantic received all of its deep water from the Southern Ocean with little vertical stratification, as observed in our stable isotope comparisons and $\epsilon_{Nd}$ isotope records (Thomas, 2005; Via and Thomas, 2006). Our stable isotope comparisons show that there was restructuring of Atlantic deep-water circulation as gateways opened and the proto-ACC deepened in the late-middle to late Eocene, with development of warm NCW and cold SCW, WSDW that may have entered the South Atlantic from the Indian Ocean, and progressive shifts in deep-water mixing. The $\delta^{13}C$ records show further complexity that indicates modification of deep-water masses by productivity.

Our isotope comparisons indicate the development of deep-water circulation patterns similar to modern AMOC, supporting models that show how early AMOC could have been driven by proto-ACC flow (Toggweiler and Samuels, 1995). The combined openings of the Drake and Tasman passages may have enhanced wind-driven overturning circulation, which may have been an early analog of modern AMOC (Toggweiler and Björnsson, 2000; Sijp et al., 2011; Fyke et al., 2015). This is analogous to the modern AMOC driven by the ACC in the South Atlantic Ocean (Garzoli and Matano, 2011).

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