

Coupling of marine and continental oxygen isotope records during the Eocene-Oligocene transition

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ABSTRACT

While marine records of the Eocene-Oligocene transition (EOT) indicate a generally coherent response to global cooling and the growth of continental ice on Antarctica, continental records indicate substantial spatial variability. Marine EOT records are marked by a $\sim +1.1\text{‰}$ foraminiferal $\delta^{18}\text{O}$ shift, but continental records rarely record the same geochemical signature, making both correlation and linking of causal mechanisms between marine and continental records challenging. Here, a new high-resolution continental $\delta^{18}\text{O}$ record, derived from the freshwater gill-breathing gastropod *Viviparus lentus*, is presented from the Hampshire Basin (UK). The Solent Group records marine incursions and has an established magnetostratigraphy, making it possible to correlate the succession directly with marine records. The *V. lentus* $\delta^{18}\text{O}$ record indicates a penecontemporaneous, higher magnitude shift ($> +1.4\text{‰}$) than marine records, which reflects both cooling and a source moisture compositional shift consistent with the growth of Antarctic ice. When combined with “clumped” isotope measurements from the same succession, about half of the isotopic shift can be attributed to cooling and about half to source moisture change, proportions similar to marine foraminiferal records. Thus, the new record indicates strong hydrological cycle connections between marine and marginal continental environments during the EOT not observed in continental interior records.

Keywords: Eocene – Oligocene transition; oxygen isotopes; climate; *Viviparus*; gastropod; continental; glaciation

1. INTRODUCTION

The Eocene-Oligocene transition (EOT), onset at ~34 Ma ago, represents a climatic regime change from "greenhouse" conditions to "icehouse" conditions. Physical evidence (various; e.g., Davis et al., 2012) indicates the initiation of large-scale Antarctic glaciation during the EOT, with permanent ice sheets formed on Antarctica for the first time between 34.0 and 33.65 Ma peaking at the onset of an oxygen isotopic event known as Oi-1 (Miller et al., 1991; Zachos et al., 1996, 2001; Coxall et al., 2005; Pälike et al., 2006, supplement), which is a ~+1.1 shift that reflects a combination of temperature and salinity/ice volume change (Zachos et al., 2008). Both benthic (Coxall et al., 2005; Coxall and Wilson, 2011) and planktonic (Pearson et al., 2008) foraminifera record a coherent spatial and temporal isotopic response, which implies global-scale changes to the Earth's oceans and global-scale climatic cooling.

In contrast, continental records are spatially variable (Sheldon, 2009; Zanazzi et al., 2015), with regional differences in the magnitude of temperature change (e.g., Zanazzi et al., 2007 vs. Retallack, 2007), the magnitude of precipitation change (Sheldon and Retallack, 2004; Abels et al., 2011), and even whether precipitation increased or decreased (Sheldon et al., 2009). The widely recognized positive isotope excursion seen in marine foraminiferal $\delta^{18}\text{O}$ records of the EOT glaciation is rarely found by terrestrial records (cf. Zanazzi et al., 2007; 2008), which instead often show no discernible shift in $\delta^{18}\text{O}$ (e.g., Sheldon et al., 2012) or a delayed shift relative to marine records (Zanazzi et al., 2007). In addition, it can be difficult to link marine and terrestrial records of the EOT for two reasons: 1) continental successions are most often preserved in endorheic basins, far from marine incursions that would make direct age comparison possible, and 2) oxygen isotope records from

continental interiors can be complicated by a variety of non-temperature related factors (e.g., changing circulation patterns, orographic effects) or may respond to local, rather than global hydrologic cycle drivers (e.g., Sheldon et al., 2012). Thus, sections in continental strata that span the EOT and which are well-calibrated to the geomagnetic polarity timescale and to marine geochronology are rare.

The English Hampshire Basin Solent Group (Fig. 1) was deposited in a coastal floodplain environment. It has documented magnetostratigraphy, sequence stratigraphy, mammalian and charophyte biostratigraphy, and brief calcareous nannoplankton events (Hooker, 1987, 2010; Sille et al., 2004; Gale et al., 2006; Hooker et al., 2009) that allow good calibration to marine records through the entire late Eocene (Priabonian) and early Oligocene (Rupelian). The depositional rate of the Solent Group strata is high, ranging from ca. 3–10 cm ky⁻¹ (Hooker et al., 2009), which facilitates high resolution sampling. The only significant hiatus is during the glacial maximum following the Oi-1 event, caused by major sea-level fall. Hren et al. (2013) recently published a “clumped isotope” paleotemperature reconstruction of the EOT from the Solent Group based on the prosobranch gastropod *Viviparus lentus* (Solander) where they found: 1) the magnitude of mean annual air temperature change (~4–6° C) was comparable to North Atlantic sea surface temperature changes, and 2) growing season temperatures for *V. lentus* were essentially constant prior to the EOT, but dropped by nearly 10° C following Oi-1. An earlier climate study of the Solent Group from the combined proxies of mammal tooth enamel, pulmonate gastropod (*Lymnaea longiscata* Brongniart) shells, fish otoliths and charophyte gyrogonites at six horizons (Grimes et al. 2005) also found little evidence for variability in summer temperatures prior to the EOT. However, both of those previous studies had limited data for the early Oligocene.

To build upon these low resolution isotope-derived climate studies and to provide greater data coverage for the early Oligocene, we have analyzed the $\delta^{18}\text{O}$ composition of shell carbonate of *V. lentus* at numerous horizons through three million years (35.4–32.4Ma) of the Solent Group (Bugler, 2011). This new, denser paleoclimatic record (with ca. 750 ky leading up to Oi-1 having a frequency interval of ca. 35 ky) is then compared with marine oxygen isotope records and with paleoclimatic records from other continental EOT sites (Fig. 2).

2. SAMPLE LOCATIONS AND METHODOLOGY

Six localities representing different parts of the late Eocene (Priabonian) to earliest Oligocene (Rupelian) Solent Group stratigraphy of the Hampshire Basin (Fig. 1; Supplemental Data and Table S1) were examined. In total, 40 stratigraphic horizons containing abundant shells of the freshwater prosobranch (gill-breathing) gastropod *V. lentus* were sampled, including both complete and fragmentary, but identifiable, shells. Modern *Viviparus* gastropods inhabit only clean, slowly moving or stable water bodies, and their habitat tracks the availability of aquatic environments (Boss, 1978). Fossil *Viviparus* gastropods have been used previously in studies of both warming (Schmitz and Andreasson, 2001) and cooling (Hren et al., 2013) Paleogene paleoclimate events, and modern *Viviparus* gastropod species reliably record changing environmental conditions (Bugler, 2011).

Bulk sediment samples were air dried, disaggregated in warm water, and then wet sieved using mesh sizes of 2 mm, 1 mm, and 250 μm . Each size fraction was oven dried at 25 °C. Between 7 and 15 randomly selected *V. lentus* fragments were picked from the >2 mm size fraction, and cleaned in an ultrasonic bath. Each individual fragment (n = 483) was crushed using an agate mortar and pestle.

Between 0.30 mg and 0.50 mg of powder was placed into vials for isotope analysis. The remainder was retained for repeat analyses and X-Ray Diffraction (XRD) in order to check for any alteration of the original aragonite mineralogy.

To determine the preservation state of shells from individual horizons, powder from 3 to 5 *V. lentus* fragments from each horizon was analyzed either at the Plymouth University (PU) or at Royal Holloway, University of London (RHUL). At PU the samples were analyzed by MJB, using a Phillips PW1792 X-ray diffractometer (XRD) with High-Score Plus identification software. The *V. lentus* powder was aligned along a metal plate, which was then placed into a sealed chamber within the XRD. A Cu anode source was used with generator settings of 30kV and 40 nA. Expert High Score was used to analyze the x-ray chromatography produced. At RHUL, samples were analyzed courtesy of Dr. Dave Alderton using a Philips Analytical XRD PW3710 machine with PC-APD diffraction software. Samples were scanned between 20° to 50° (2θ) using a copper tube anode.

For isotope analysis the carbonate powders were reacted with 100% phosphoric acid at 90 °C for approximately 1 hour. The CO₂ produced was analyzed on an Isoprime Mass Spectrometer with a Gilson Multiflow carbonate auto-sampler at Plymouth University. The results were calibrated against Vienna Pee Dee Belemnite (VPDB) using the international standards NBS-19, IAEA -CO-8, and IAEA-CO-9. Five NBS-19 standards were also evenly distributed throughout the individual isotope runs to correct for daily drift. The mean standard deviation on replicate analyses of individual samples was on the order of 0.2‰ for δ¹³C and 0.2‰ for δ¹⁸O. Analytical results are given in Supplemental Table S1.

The lithostratigraphy used here (Figs. 1,3-4) follows Hooker et al. (2009). Calibration of the Solent Group to the geomagnetic polarity scale (Vandenberghe et

al. 2012) also follows Hooker et al. (2009), but the timescale used is that of Pălike et al. (2006, supplement). Calibration is based on combined biostratigraphy, magnetostratigraphy and sequence stratigraphy, allowing for correlation between both terrestrial and marine records. The highest resolution has been obtained where biostratigraphic events coincide with either sequence boundaries or magnetic polarity shifts, or both. Thus, in terms of samples used in this study, the most accurately positioned points are at the boundaries of normal and reversed polarity zones (where these are recorded), the base of the Lacey's Farm Member, the base of the Bembridge Limestone Formation, and top of the lower Hamstead Member. The hiatus during the Oi-1 glacial maximum is estimated as being the interval (Fig. 3) between the top log bed (tlb) sequence boundary and the subsequent maximum flooding surface of the *Nematura* Bed (i.e., basal meter of the upper Hamstead Member; Hooker et al. 2009), identified as belonging to sequence TA4.4 of Haq et al. (1987). Between these points, resolution is less precise and assumes constant sedimentation rate. Sampling from the base of the Seagrove Bay Member to the top of the lower Hamstead Member, representing ca. 750 ky (EOT) leading up to Oi-1, averages a frequency interval of ca. 35 ky. The sampling frequency in the last ca. 250 ky subset averages ca. 18 ky.

3. RESULTS

X-ray diffraction of 173 *V. lentus* fossil fragments from 40 horizons (22 spanning the EOT and Oi-1) indicates preservation of aragonite consistent with that of another freshwater gastropod, *L. longiscata*, from the same clay-dominated lacustrine strata (Grimes et al., 2005). This, in addition to the preservation of three discrete carbonate layers, as viewed in numerous cross sections through

representative shells (Bugler, 2011; Hren et al., 2013), strongly suggests that the *V. lentus* fossil fragments are not diagenetically recrystallized and therefore should preserve their original isotopic signal. These results are also supported by previous high-resolution single-shell microsampling that indicated that the *V. lentus* shells of the Solent Group preserve sub-annual seasonal variability in $\delta^{18}\text{O}$ (Hren et al., 2013).

The $\delta^{18}\text{O}$ record from *V. lentus* is shown in Figures 3-4. At each of the 40 stratigraphic levels the $\delta^{18}\text{O}$ data points are a mean from the analyses of at least 7 *V. lentus* shell fragments and the associated errors (shaded region) are the standard deviation ($\pm 1 \sigma$) from the mean. Bugler et al. (2009), based upon micro-milled isotope profiles of a number of modern *Viviparus contectus* (Millet) shells, demonstrated that $\delta^{18}\text{O}$ values could vary cyclically by up to 1.9‰, primarily as a response to changes in seasonal temperature. Similar magnitude $\delta^{18}\text{O}$ cycles were also shown by Hren et al. (2013) in a smaller number of whole *V. lentus* shells from the Solent Group, which suggests that the variability reported here from the analysis of multiple shell fragments is likely to be related to seasonal changes.

Early in the EOT, $\delta^{18}\text{O}$ values from *V. lentus* shells are variable, but generally cluster around -1.5‰ (Figs. 3a, 4a). Just prior to Oi-1 (upper half of the lower Hamstead Member), they shift to -0.6 then back to -1.7‰ at 33.7 Ma. At Oi-1, $\delta^{18}\text{O}$ values shift to + 0.1‰ (basal upper Hamstead Member) before a gradual recovery toward more depleted values. The timings of these shifts are broadly similar to marine isotope records, with the Solent Group excursions and recoveries each roughly 0.5‰ larger than the marine isotope records. The pattern of a minor followed by a major positive shift resembles the succession EOT-1 (precursor event) and Oi-1 in marine sites (e.g., Coxall et al., 2005; Coxall and Wilson, 2011; herein, Fig. 3b).

4. DISCUSSION

4.1. Suitability of using *V. lentus* as a continental oxygen isotope proxy

Viviparus lentus, which has been used successfully before in paleoclimate research (Schmitz and Andreasson, 2001; Hren et al., 2013), is an ideal freshwater paleoclimate proxy because modern *Viviparus* colonizes a wide range of habitats including rivers, streams, ponds and lakes (Boss, 1978). They are also gill breathers (prosobranchs) and therefore, are found in permanent water bodies up to 20 m deep (Boss, 1978). Their dependence on dissolved oxygen for respiration makes them intolerant of polluted water (Strayer, 1990), so they are rarely found in stagnant water bodies (e.g., shallow ponds; Jokinen, 1983), where significant evaporation can occur. These ecological parameters, in addition to evidence for the maintenance of primary aragonite and a 3 layered shell structure, all suggest that the isotope record that has been generated from *V. lentus* is unaffected by diagenesis or evaporative enrichment.

4.2. Factors controlling shifts in the *V. lentus* $\delta^{18}\text{O}$ record

The magnitude of the $\delta^{18}\text{O}$ shifts differs significantly between the continental *V. lentus* and the marine foraminiferal records. The marine $\delta^{18}\text{O}$ shifts are typically of the order 0.4–1.0‰, while those from the Hampshire Basin are >1.4‰ (Fig. 3). Oxygen isotope values observed in the marine realm represent a combination of temperature and the oxygen isotopic composition of seawater, which is controlled mainly by changes in the global ice volume.

In the marine realm, a well-mixed ocean is assumed (various, e.g., Zachos et al., 2001) and consequently, shifts in the oxygen isotopic composition of seawater are related largely or entirely to ice volume and temperature changes. While

temperature plays a role in the continental isotope record, what controls the isotopic composition of the water source from which *V. lentus* precipitates is more complicated. These factors include (i) evaporation, (ii) changes in the isotopic ratio of precipitation, and (iii) change in sources of precipitation. These additional factors, the last two of which may be linked to changes in global ice volume, probably explain the greater magnitude of the isotope shifts seen in the continental realm.

In section 4.1 an ecological argument against evaporation playing a dominant role in controlling the *V. lentus* $\delta^{18}\text{O}$ record was presented. In other continental EOT successions (Fig. 2), locally evaporative conditions are indicated by aridification (Sheldon and Retallack, 2004; Abels et al., 2011; Passchier et al., 2013; Boardman and Secord, 2013), by shifts in the assemblage of paleosol types (Terry, 2001), and by shifts in the trace fossil assemblage (Sheldon and Hamer, 2010). None of these features are observed in the Hampshire Basin, and rather than aridification through the EOT, wetter conditions are reconstructed (Fig. 4c; Sheldon et al., 2009) using the same reconstruction technique that indicated aridification in other regions (see Supplemental Table S2). Furthermore, the most pronounced evidence for evaporative enrichment in the Hampshire Basin is at ca. 34.4 Ma ago, based on $\delta^{18}\text{O}_{\text{water}}$ values calculated using clumped isotope data (Hren et al., 2013). In addition, some of the paleosols in the Hampshire Basin stratigraphic sequence also include various redoximorphic features such as drab-haloed root traces and rare Fe-Mn nodules that are consistent with episodic or even continuous inundation (Sheldon et al., 2009). Thus, in contrast to some other continental records that are potentially compromised by local evaporative conditions (e.g., Nebraska; Sheldon, 2009; Boardman and Secord, 2013; c.f. Zanazzi et al., 2015), the Hampshire Basin record appears likely to be a high-fidelity recorder of environmental conditions.

At the same time, if the $\delta^{18}\text{O}$ shifts (Fig. 3a, 4a) are due to temperature change alone, then they are above, or at the very upper limit of, those previously recorded. For example, using the *Viviparus* genus specific equation (Bugler et al. 2009), and assuming no change in the $\delta^{18}\text{O}$ of freshwater, temperature shifts across the EOT would be between 10 and 13°C. These reduce to 6–8°C using a general inorganic aragonite equation (Grossman and Ku, 1986) or if the likely season of the majority of carbonate formation is used to adjust the temperature (Hren and Sheldon, 2012). In contrast, a sea surface temperature (SST) change of ~5°C across the EOT has been reconstructed recently based on high latitude marine records of TEX_{86} and $\text{U}^{\text{K}'}_{37}$ (Liu et al. 2009). Similarly, Hren et al. (2013) documented a 4–6°C drop in mean annual air temperature across the EOT (Fig. 4b) using *V. lentus* as in this study, and “clumped isotope” analyses, which are insensitive to evaporation or to changes in source water $\delta^{18}\text{O}$. Thus, the discrepancy between the apparent magnitude of temperature change indicated by the new $\delta^{18}\text{O}$ record if due to temperature alone (6–13°C depending on equation) and the “clumped isotope” record (4–6°C) likely indicates a shift in source water $\delta^{18}\text{O}$ toward relatively enriched values, which accompanied the drop in temperature. Because source water evaporative enrichment was unlikely to have been significant (see above), the shift in source water $\delta^{18}\text{O}$ toward enriched values instead likely records enrichment in meteoric water due to ice sheet growth on Antarctica. Assuming that the “clumped isotope” record (Hren et al., 2013) and high-latitude marine SST (Liu et al., 2009) reflects the “true” temperature change, then roughly half of the $\delta^{18}\text{O}$ shift recorded in the new Hampshire Basin *V. lentus* record is due to temperature change and half is due to ice volume change. These relative proportions are similar to what has been inferred for marine foraminiferal $\delta^{18}\text{O}$ records (Zachos et al., 2001; Coxall et al., 2005). In addition, the

marine SST (Liu et al., 2009) and “clumped isotope” records (Fig. 4b) both indicate a post-EOT rebound to slightly warmer conditions that is mirrored by marine $\delta^{18}\text{O}$ records and by the Hampshire Basin *V. lentus* $\delta^{18}\text{O}$ record (Figs. 3a, 4a), and by reconstructed atmospheric $p\text{CO}_2$ based on the $\delta^{11}\text{B}$ composition of planktonic foraminifera (Pearson et al., 2009).

Additional support for interpretation of the enriched $\delta^{18}\text{O}$ values during the EOT and Oi-1 as representing both ice-volume and temperature is provided by sequence stratigraphy. The highest $\delta^{18}\text{O}$ values and lowest clumped isotope temperatures postdate the top log bed (tlb) sequence boundary (Fig. 3), where an unconformity indicates major sea-level fall. In contrast, lower in the succession pre-EOT (Fig. 3a), high $\delta^{18}\text{O}$ values coincide with sequence boundaries (Lacey’s Farm and Seagrove Bay Members), while shifts towards lower values coincide with maximum flooding surfaces (Fishbourne and Bembridge Marls Members), where temperatures recorded by clumped isotopes are relatively stable (Fig. 4b). This suggests that changes in ice-volume, not temperature, were responsible for the $\delta^{18}\text{O}$ peaks and troughs during the ca. 1my preceding the EOT.

Observations of modern *V. contectus* in England indicate that shell growth is dominated by summer growth (Bugler et al., 2009). Ecology of *V. lentus* preserved in the Hampshire Basin was likely similar, even if the overall climate state was warmer during the EOT. Lake water temperatures are very highly correlated with air temperatures, but the precise relationship varies quantitatively according to latitude and also to the length of season that is considered (Hren and Sheldon, 2012). Thus, while it is difficult to ascertain exactly what part of the year is represented by the Hampshire Basin record, it is likely most reflective of summer growing season lake conditions. Serial isotope sampling of *V. lentus* from the Hampshire Basin indicates

similar season ranges before, during, and after the EOT, but with different absolute values during EOT, suggesting substantial reorganization of the hydrologic cycle (Bugler, 2011; Hren et al., 2013). While the large sustained $\delta^{18}\text{O}$ shift recorded in the Hampshire Basin is not recorded in European endorheic basin records (Maiaans, Spain; Fig. 4e), similar reorganization of the hydrologic cycle was previously identified from high magnitude variability in paleosol carbonate $\delta^{18}\text{O}$ values (Sheldon et al., 2012). Because the poles cooled more rapidly during the EOT than lower latitude sites (e.g., Liu et al., 2009), the changing ocean and air temperature gradients may have changed the moisture source amounts more at higher latitudes as well. Although records from other sites are needed to confirm this pattern, it suggests that ice sheet growth on Antarctica was connected to Northern Hemisphere climate through cooling of North Atlantic ocean waters and alteration of regional hydrology.

4.3 Comparison with other terrestrial paleoclimate records

As discussed in the *Introduction*, terrestrial records of the EOT indicate varied responses and varied magnitudes of response (Supplemental Table S2; Fig. 4). For example, records from the Hampshire Basin (this study; Hren et al., 2013), Nebraska (Zanazzi et al., 2007; but see also Boardman and Secord, 2013), and Oregon (Sheldon and Retallack, 2004) indicate significant cooling, while records from Montana (Retallack, 2007) and Spain (Sheldon et al., 2012) indicate little or no cooling. There are no systematic differences between the type of proxy employed (i.e., paleosol records can indicate either cooling or stability depending on the locality), which suggests that this environmental heterogeneity is real. Summarizing from various studies, two general trends have emerged: (1) proximity to a marine

moisture source matters, and (2) the magnitude of climatic response is larger at higher paleolatitudes. In comparing a number of paleoprecipitation reconstructions, Sheldon et al. (2012) recognized that records from endorheic basins indicated little response whereas basins that were exorheic and located close to their marine moisture source (e.g., Hampshire Basin) recorded a strong response. This is similar to the observation that modern coastal precipitation stations record $\delta^{18}\text{O}$ of precipitation that more closely mirrors changes in surface ocean water $\delta^{18}\text{O}$ than inland sites, due to continentality effects (Dansgaard, 1964). Similarly, terrestrial temperature response to the EOT (Fig. 4) appears to reflect paleolatitude, where higher paleolatitude sites (Fig. 2; e.g., Hampshire Basin; Blakey, 2008) indicate higher magnitude cooling than relatively lower paleolatitude sites (e.g., Oregon), a pattern which mimics marine SST records (Liu et al., 2009). Thus, when combining both patterns, a mid-latitude endorheic basin site such as the Ebro Basin (Fig. 4e; Sheldon et al., 2012) shows little or no climatic response to the EOT, whereas a higher latitude coastal plain site such as the Isle of Wight (Fig 4a-b; *this study*) records the highest magnitude changes, larger even than other sites located relatively close to the ocean (e.g., Oregon). In Oregon, the Cascades uplift post-dated the EOT and causes a rainshadow at present that was not a factor in the Paleogene. Furthermore, accretion of the Coast Range generally post-dates the Eocene, so central Oregon was much closer to the coastline at the time of deposition than it is at present (Fig. 2).

5. CONCLUSIONS

XRD data, shell structural preservation, and ecology of nearest living relatives suggest that *V. lentus* from the UK Solent Group yields a $\delta^{18}\text{O}$ record that has not

been affected by diagenesis. Based on paleosol proxy evidence, precipitation increases up sequence and does not shift in amount in parallel with the *V. lentus* $\delta^{18}\text{O}$ record, further indicating that evaporation is unlikely to be a major factor affecting these continental oxygen isotope values in the Solent Group during the EOT.

The oxygen isotope record derived from *V. lentus* records a positive $\delta^{18}\text{O}$ shift across Oi-1 of $>1.4\text{‰}$ that is roughly coincident with marine foraminiferal records. If entirely due to temperature, this would equate to a high magnitude cooling, contrasting with marine SST from roughly the same paleolatitude and continental “clumped isotope” records from the same site. This implies a contribution to the $\delta^{18}\text{O}$ shift from another factor, such as changes in the isotopic composition of the meteoric water in which *V. lentus* grew. The most likely cause of this change, supported by coincidence with sea-level fall, is the growth of ice sheets on Antarctica, which would have shifted marine source waters toward more enriched $\delta^{18}\text{O}$ values. Thus, while the magnitude of the total $\delta^{18}\text{O}$ shift recorded in the Hampshire Basin is larger than in marine records, the relative contribution of temperature and ice volume changes appears to be the same. This, coupled with comparison of continental sites that are more distal relative to marine moisture sources and at other paleolatitudes, indicates strong terrestrial-marine coupling during the EOT via the global hydrological cycle and provides a possible explanation for the observed spatial heterogeneity of terrestrial EOT responses.

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524

FIGURE CAPTIONS

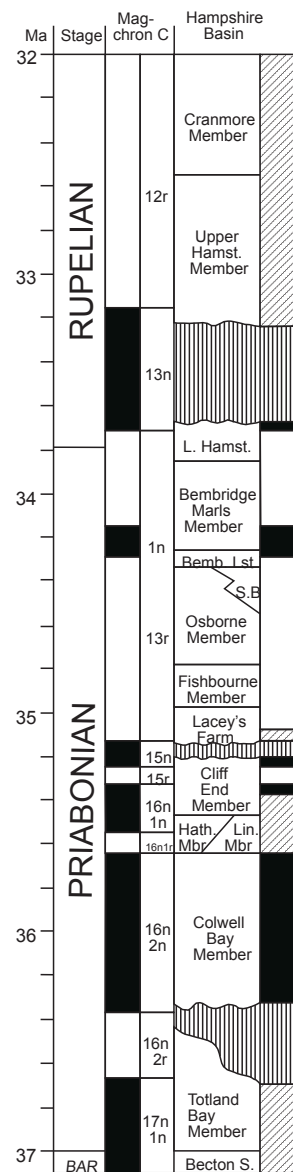
Figure 1. Sample locations and their stratigraphic context .The extent of sections selected for sampling at sites A-F is shown against the stratigraphic column of Hooker et al. (2009). Map modified from Daley (1999). The timescale used here and in Figures 3-4 is that of Pälike et al. (2006, supplement). Stratigraphic abbreviations: BAR = Bartonian; Becton S. = Becton Sand Formation; Bembridge Lst = Bembridge Limestone Formation; Hamst. = Hamstead; Hath. = Hatherwood Limestone; L. = lower; Lin. = Linstone Chine; Mbr = Member; S.B. = Seagrove Bay Member.

Figure 2. Locations of marine and terrestrial Eocene-Oligocene transition sites. Marine comparative sites include Ocean Drilling Project site 1218 (IODP; e.g., Coxall et al., 2005) and from the Tanzanian Drilling Project sites 12 and 17 (TDP; e.g., Pearson et al., 2008). North America terrestrial sites are reviewed in Sheldon (2009) and the Ebro Basin (Spain) site is described in Sheldon et al. (2012). Late Eocene paleogeographic base map is used with permission from Ron Blakey, Colorado Plateau Geosystems (original reference: Blakey, 2008).

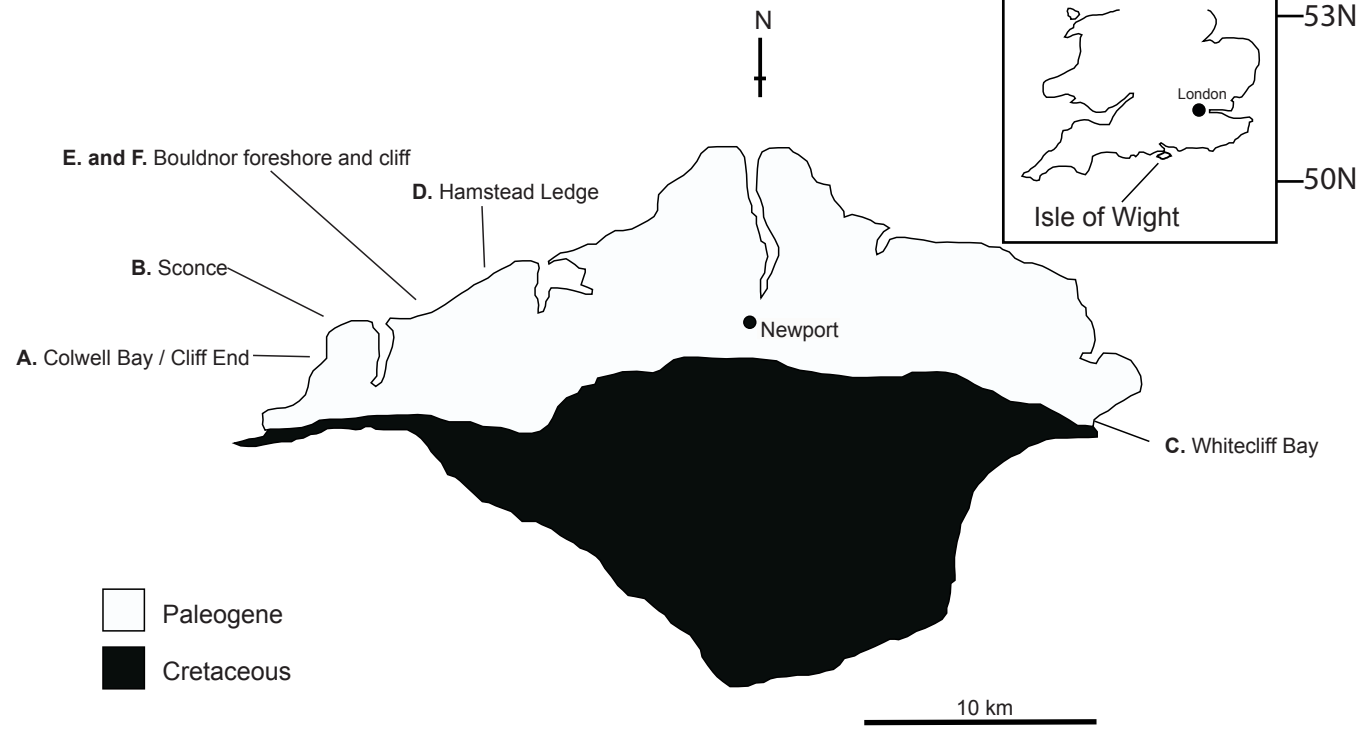
Figure 3. Comparative climate data for the Solent Group and marine records. a) The mean $\delta^{18}\text{O}$ record from *V. lentus* shell fragments plotted against the Solent Group stratigraphy (Hooker et al., 2009). Shaded area around the mean $\delta^{18}\text{O}$ values represents $\pm 1 \sigma$ (standard deviation) from the mean. b) Isotope results from the Ocean Drilling Program (ODP) site 1218 (Coxall et al., 2005; see also Coxall and Wilson, 2011) and from the Tanzanian Drilling Project sites 12 and 17 (Pearson et al. 2008; Lear et al. 2008). Abbreviations: mfs = maximum flooding surface (marked by

maximum salinity fauna within sequence); sb = sequence boundary; tlb = top log bed;
EOT = Eocene-Oligocene transition; others as in Fig. 1.

Figure 4. Comparative climate data for the Solent Group and other terrestrial sites. Panels (a)-(c) are for data from the Solent Group and panels (d) and (e) are for the John Day Basin (Oregon) and the Ebro Basin (Spain), respectively. a) *V. lentus* $\delta^{18}\text{O}$ record as in Figure 3. Shaded area around the mean $\delta^{18}\text{O}$ values represents $\pm 1\sigma$. b) Summer season temperatures calculated by Grimes et al. (2005) and mean annual air temperatures calculated by Hren et al. (2013) using the April to October water-air temperature transfer function of Hren and Sheldon (2012). c) Paleosol-derived mean annual precipitations calculated for the Solent Group by Sheldon et al. (2009). d) Paleosol-derived mean annual air temperature calculated for the John Day Basin by Sheldon (2009). e) Paleosol-derived mean annual air temperature calculated for the Ebro Basin by Sheldon et al. (2012). Representative error bars for (d; $\pm 4^\circ\text{C}$) and (e; $\pm 2.5^\circ\text{C}$) are given for the oldest paleosol and are the same for all the other data points, but have not been plotted throughout for clarity's sake. While the records from near coastal (Solent Group) and continental margin (Oregon) settings show cooling associated with the EOT and a general temporal correspondence with the marine realm, intermontane continental interior sites (e.g., Spain) show relatively little change during the EOT. Abbreviations: MAP = mean annual precipitation; MAAT = mean annual air temperature; others as in Fig. 1.



F
E
D
C
B
A



□ Paleogene
■ Cretaceous

