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# Modelling the effects of sediment compaction on salt marsh reconstructions of recent sea-level rise

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ABSTRACT

This paper quantifies the potential influence of sediment compaction on the magnitude of nineteenth and twentieth century sea-level rise, as reconstructed from salt marsh sediments. We firstly develop a database of the physical and compression properties of low energy intertidal and salt marsh sediments. Key compression parameters are controlled by organic content (loss on ignition), though compressibility is modulated by local-scale processes, notably the potential for desiccation of sediments. Using this database and standard geotechnical theory, we use a numerical modelling approach to generate and subsequently 'decompact' a range of idealised intertidal stratigraphies. We find that compression can significantly contribute to reconstructed accelerations in recent sea level, notably in transgressive stratigraphies. The magnitude of this effect can be sufficient to add between 0.1 and 0.4 mm $yr^{-1}$ of local sea-level rise, depending on the thickness of the stratigraphic column. In contrast, records from shallow ($<0.5$ m) uniform-lithology stratigraphies, or shallow near-surface salt marsh deposits in regressive successions, experience negligible compaction. Spatial variations in compression could be interpreted as 'sea-level fingerprints' that might, in turn, be wrongly attributed to oceanic or cryospheric processes. However, consideration of existing sea-level records suggests that this is not the case and that compaction cannot be invoked as the sole cause of recent accelerations in sea level inferred from salt marsh sediments.



## 1. Introduction and aim

Considerable interest is focussed on the possibility of an acceleration in global sea-level rise that has occurred in the last 100–200 years. Twentieth century rates of global sea-level rise recorded by tide gauges (1.7–1.8 mm $yr^{-1}$; Bindoff et al., 2007; Church and White, 2006; Douglas, 1991) are greater than those obtained from the geological record of Late Holocene sea-level rise (e.g. Shennan and Horton, 2002; Engelhart et al., 2009; Gehrels et al., 2012). Constraining the magnitude, timing and spatial variability of this acceleration is critical to our understanding of atmosphere–cryosphere–ocean interactions. However, there are few pre-1900 tide gauge records and those that exist are concentrated in NW Europe, limiting their use in resolving this matter (Douglas, 1992; Woodworth et al., 2009). An alternative approach is to use the proxy sea-level records contained within salt marsh stratigraphies (Gehrels, 2000; Scott and Medioli, 1978). Consequently, an increase in the rate of sea-level rise in the late nineteenth to early twentieth century has been identified in several sea-level reconstructions derived from salt marshes in the northern (Donnelly et al., 2004; Gehrels et al., 2005, 2006; Kemp et al., 2009; Leorri et al., 2008) and southern hemispheres (Gehrels et al., 2008, 2012). This phenomenon has been attributed to a combination of natural and anthropogenic influences (cf. Kemp et al., 2011; Mann and Jones, 2003).

A critical assumption of the salt marsh approach is that the effects of sediment compaction on these low-energy, organic sedimentary deposits are negligible. Compaction involves a combination of physical and biochemical processes that reduces the vertical thickness of the sediment column (Allen, 2000), distorting stratigraphic sequences (Allen, 1999; Bloom, 1964; Haslett et al., 1998; Jelgersma, 1961; Kaye and Barghoorn, 1964). Compaction lowers 'sea-level index points' (SLIs) from their depositional altitudes (Edwards, 2006), introducing a potentially significant error into sea-level reconstructions obtained from salt marshes and associated low energy intertidal stratigraphies (Long et al., 2006). Since the effects of compaction are time-dependent, there is a further possibility that surface lowering of a salt marsh could be misinterpreted as a 'real' acceleration in the rate of sea-level rise. Unfortunately, the contribution of compaction to

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salt marsh-based reconstructions of recent sea level is currently poorly constrained. Our aim is to address this critical issue by quantifying the potential contribution of sediment compression, a key compaction process that describes volumetric changes in response to applied overburden pressures, to the recent acceleration in salt marsh-based sea-level reconstructions.

## 2. Sediment compaction

Late Holocene stratigraphic studies report rates of compaction that are regional estimates averaged over centennial to millennial timescales and/or obtained from multiple, often thick (5–10 m) stratigraphic sections (Horton and Shennan, 2009; Törnqvist et al., 2008). These studies lack sufficient resolution to be applicable to the shallower sections ( < 3 m) and shorter timescales (decadal to centennial) typical of recent salt marsh studies. Basal salt marsh peats that directly overlie an incompressible surface can provide a compaction-free sea-level record (Törnqvist et al., 2004). However, to date only one (Donnelly et al., 2004) of the published high-resolution salt marsh reconstructions of sea level for the last 200 years uses basal peats; the others rely on continuous salt marsh stratigraphies.

Numerical modelling of compaction provides an alternative approach to solely field-based studies (Massey et al., 2006; Paul and Barras, 1998; Pizzuto and Schwendt, 1997). However, such models often employ classical soil mechanics theories that may not be suitable in low energy intertidal settings, where unique lithologies form and where the stress and diagenetic environments are distinct. Indeed, Brain et al. (2011) studied the compression behaviour of minerogenic tidal flat and salt marsh sediments and found a difference to that assumed in standard geotechnical compression models because of overconsolidation at the depositional surface, meaning that they have experienced an effective compressive stress greater than that exerted by the existing overburden (Powrie, 2004). Overconsolidated sediments exhibit a reduced compressibility state until the previous maximum value of effective stress experienced by the sediment (the yield stress) is exceeded. Brain et al. (2011) explained the observed compression behaviour of these sediments in terms of organic content and the frequency and duration of tidal immersion, which are both controlled by elevation within the intertidal zone, and developed a theoretical framework that describes compression behaviour within NW European low energy intertidal settings. This framework describes variations in compression behaviour with reference to a four-parameter model (Fig. 1).

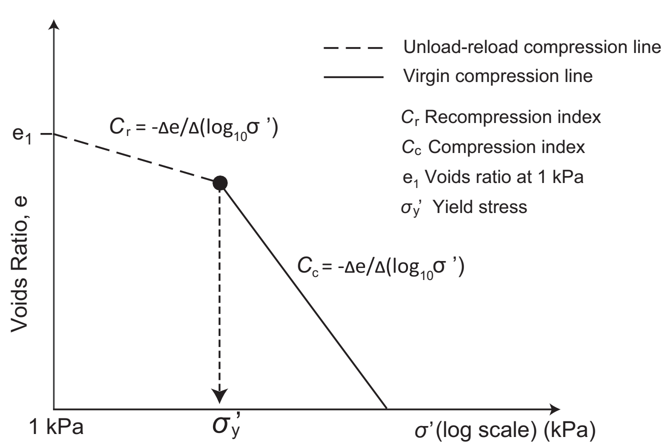


**Fig. 1.** Four-parameter model to describe compression behaviour in low energy intertidal sediments. See text for further description.

## 3. The compression property database

### 3.1. Field data collection

We studied three UK field sites that vary in geomorphic setting, hydrographic conditions and eco-sedimentological characteristics: Cowpen Marsh, Roudsea Marsh and Thornham Marsh (Fig. 2; see Supplementary information). At each site, we recorded vegetation assemblages, logged sample lithology and described the physical and geotechnical properties of samples in accordance with Keble Martin (1974) and British Standards Institute (2002).

We collected undisturbed sediment samples (British Standards Institute, 1999) from the upper 0.2 m of the sediment column (i.e. the depositional surface) from each eco-sedimentological zone for laboratory testing. Samples were stored in confined, sealed and cold conditions (*ca.* 4 °C) to prevent disturbance due to stress relief, moisture loss and to limit bacterial processes.

### 3.2. Laboratory methods

We determined organic content by loss on ignition (LOI; 550 °C for 4 h) and particle size by laser granulometry (see Brain et al., 2011). We determined sediment physical properties (moisture content, specific gravity, bulk density, voids ratio by Height of Solids) in accordance with BS 1377 (British Standards Institute, 1990; Head, 1988). One-dimensional, $K_0$ (zero lateral strain) compression testing was undertaken using fixed ring, front-loading oedometers in general accordance with BS 1377 (British Standards Institute, 1990; Head, 1988), but with modifications outlined by Brain et al. (2011). Each loading stage lasted 24 h.

### 3.3. Numerical and analytical techniques

Due to the differences in tidal range between the sites, it is necessary to standardise water levels to allow comparison of elevation data, which we use as a surrogate for the frequency and duration of tidal flooding and subaerial exposure. We have used the Standardised Water Level Index (SWLI) method (Horton and Edwards, 2006) which linearly standardises the height between two tidal levels, Mean High Water of Spring Tides (MHWST, where SWLI = 100) and Mean Low Water of Spring Tides (MLWST, where SWLI = 0). We have estimated the compressive yield stress, $\sigma'_y$, using changepoint regression modelling (Carlin et al., 1992; Lunn et al., 2000; Parnell, 2005).

### 3.4. Compression behaviour and properties

LOI, initial voids ratio ($e_i$, the voids ratio at the depositional surface) and compression indices ($C_r$ and $C_c$) all vary directly with SWLI (Fig. 3). $e_i$, $C_r$ and $C_c$ all covary strongly with LOI, suggesting that organic content exerts a significant control on these compression properties (Fig. 4). These relationships occur despite variability in site characteristics (see Supplementary information).

Similar trends in yield stress with SWLI were observed at each of the three sites. These trends are characterised by a rise in $\sigma'_y$ from minima in the sandflat and mudflat samples to peaks in the low and mid-marsh zones. From peak values, $\sigma'_y$ decreases with SWLI to minima in the high marsh (Fig. 3). We observed some local variability in absolute values, but not in trends. We attribute reduced values of $\sigma'_y$ at Roudsea Marsh to the sand-rich nature of the intertidal zone, which prevents high suction stresses from being achieved during groundwater falls and subaerial desiccation.

### 3.5. Controls on compression behaviour

The relationships between lithology and $e_i$, $C_r$ and $C_c$ have a sound physical basis. High organic contents reflect the growth of

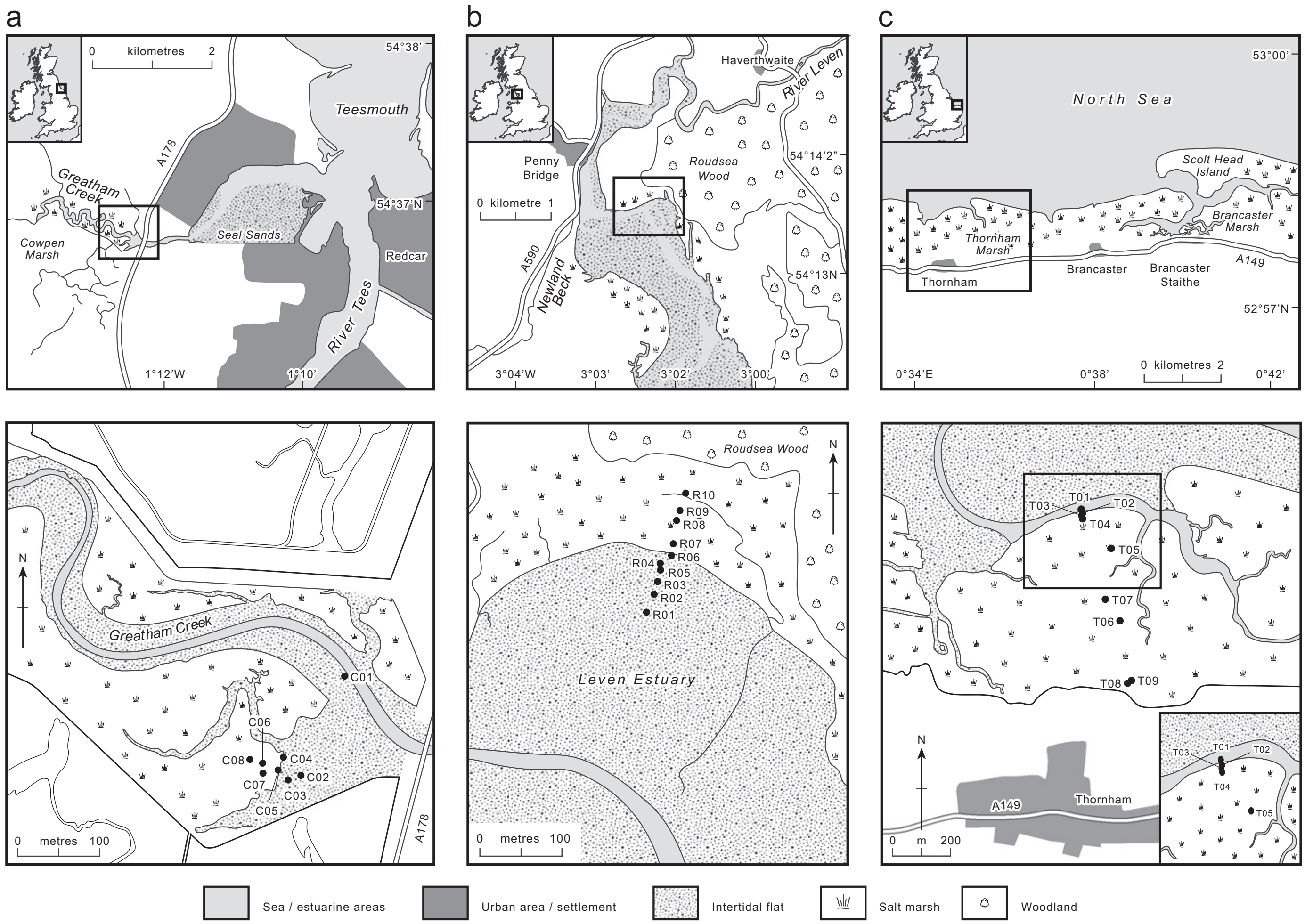


**Fig. 2.** Field site and sampling locations maps for (a) Cowpen Marsh, (b) Roudsea Marsh and (c) Thornham Marsh.

vascular plant species, which create well-aerated, highly porous soil structures (Delaune et al., 1994), and hence higher values of $e_i$. In addition, reduced flow velocities on salt marshes caused by the frictional drag of vegetation (e.g. Leonard and Luther, 1995) create more open microstructures compared to those observed in tidal flat environments, where flow velocities are greater (Burland, 1990). Less dense sediments are more prone to compression than their denser equivalents (Burland, 1990; Price et al., 2005; Skempton, 1970). In addition, organic-rich sediments are more prone to compression than their minerogenic equivalents, since organic matter is compressible (Head, 1988). In contrast, sediments with lower values of $e_i$, $C_r$ and $C_c$ accumulate lower in the intertidal zone (Fig. 3) where suspended sediment concentrations are greater during periods of tidal flooding (Shi et al., 2000). Hence, denser sediments (lower values of $e_i$) form here as a result of higher sedimentation rates from a denser suspension (Been and Sills, 1981; Sills, 1998). The absence of vascular plant growth also prevents the formation of organogenic, open microstructures.

We suggest that the yield stresses, $\sigma'_y$, recorded in the surface sediments result from two interacting factors. The rising trend in $\sigma'_y$ that peaks in the low marsh (Fig. 3) results from increasing subaerial desiccation associated with a reduction in flooding frequency and duration with increasing elevation. The post-peak decline in yield stress in the higher marsh, most apparent at Cowpen Marsh but also suggested by the data for Thornham and Roudsea Marshes (Fig. 3), likely records the shift from below-ground to above-ground productivity as vegetation assemblages change (de Leeuw and Buth, 1991). Increased above-ground biomass permits the production of decaying mats of vegetation, which increase in thickness with elevation (see Supplementary information) and reduce the potential for desiccation. Secondly, an increase in organic content reduces the influence of the cohesive component of the sediment in sustaining suction pressures (cf. Hawkins, 1984). Air entry and desaturation are likely to occur at significantly lower suction stresses than in the fine-grained, minerogenic sediments of the mudflat and low marsh zones, thereby limiting the maximum yield stress that can be achieved by desiccation.

### 3.6. An improved compression framework

Our findings help us to refine the compression framework for intertidal sediments proposed by Brain et al. (2011). Importantly, our new dataset confirms that values of near-surface voids ratio ($e_1$) and compression indices ($C_r$ and $C_c$) are controlled by organic content (a function of intertidal elevation), regardless of geomorphic setting and eco-sedimentary conditions (Fig. 4).

The controls on yield stress vary in response to feedbacks relating to organic content and the proportion of above-ground biomass production (again, both functions of elevation) that reduce the value of yield stress observed in the high marsh. Despite similar trends observed at each of the sites, absolute values of $\sigma'_y$ and the ways in which they vary are site-specific. Nevertheless, it is possible to fit a statistically significant regression model to the variations in $\sigma'_y$ with SWLI with sufficient data, as demonstrated in Fig. 4(d). Minor extrapolation suggests that

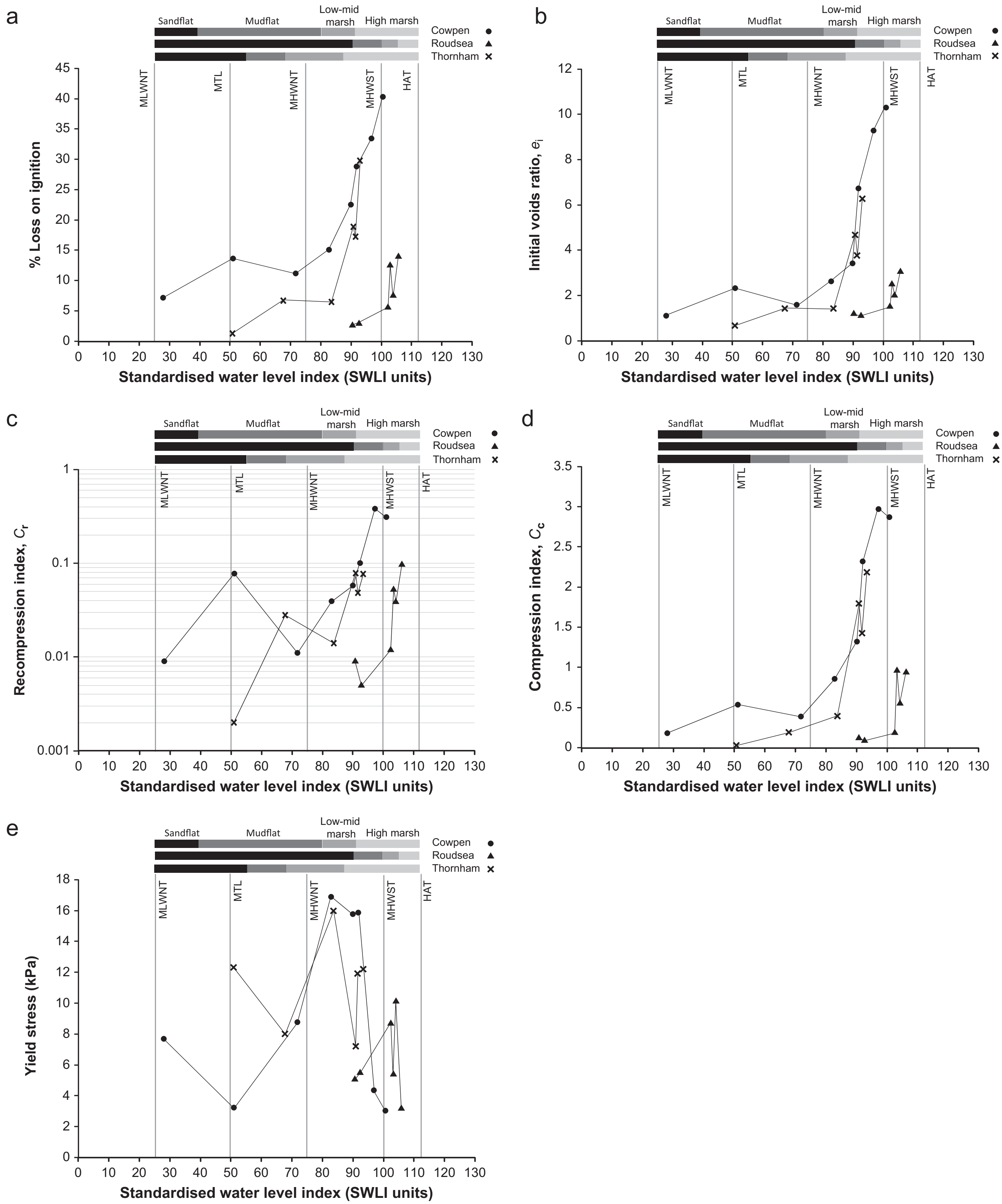


**Fig. 3.** Variations in loss on ignition (a) and compression model parameters (b, c, d, e) with standardised water level index at Cowpen, Roudsea and Thornham Marshes. MLWNT is Mean Low Water Neap Tide level; MTL is Mean Tide level; MHWNT is Mean High Water Neap Tide level; MHWST is Mean High Water Spring Tide level; HAT is Highest Astronomical Tide level.

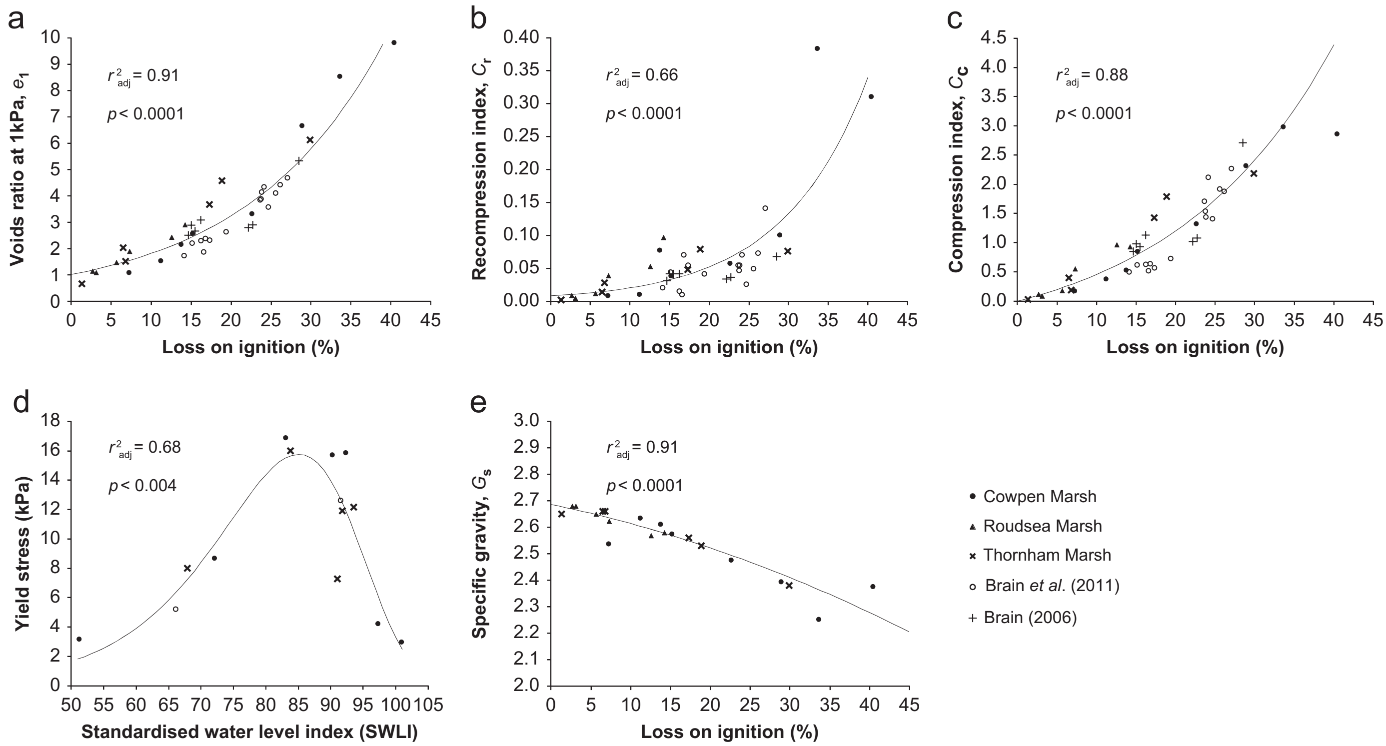


**Fig. 4.** Variations of key compression and physical properties with controlling variables, demonstrating the goodness of fit and statistical significance of predictive regression models. Where available, new data are supplemented with data from previous studies at Cowpen Marsh using contemporary sediments (Brain et al., 2011) and near-surface core samples (Brain, 2006). In (d), we use only data from Cowpen Marsh and Thornham Marsh to develop the regression model, which display similar trends. In addition, we do not use samples obtained from sandflat deposits, since these environments are not investigated as part of the modelling element.

the highly organic sediments forming in the high marsh are likely to be only very lightly overconsolidated, or indeed normally consolidated. Hence, high marsh sediments are the most compressible (high $C_r$ and $C_c$), but also may not experience a low compressibility phase (low or absence of $\sigma'_y$). However, the minimum effective stress value replicated in our oedometer tests is *ca.* 3 kPa and so we have no direct empirical evidence of fully normally consolidated sediments.

The strength and significance of the relationships in Fig. 4 indicate that the first-order, routine decompaction of intertidal sediments may be possible without the need for extensive geotechnical testing (cf. Paul and Barras, 1998; Skempton, 1944). This is potentially an extremely useful tool for sea-level research, since LOI and SWLI values are routinely determined during stratigraphic and basic laboratory investigations. However, the current database would need to be expanded to ensure that the characteristics of the study sites of interest are adequately captured (see Section 6.5).

## 4. Compression modelling

### 4.1. Modelled stratigraphic successions

We now generate, and then 'decompact', a series of synthetic stratigraphic successions that vary in thickness and the depositional sequencing of lithologies. We define stratigraphies that broadly reflect the thickness ranges of existing salt marsh records of late Holocene/twentieth century accelerations (Gehrels et al., 2005, 2006, 2008, 2012; Kemp et al., 2009; Leorri et al., 2008). Such stratigraphies range in contemporary thickness from *ca.* 0.5 to 3 m. The acceleration is generally recorded at depths $<0.5$ m below ground level (m bgl).

The ideal lithostratigraphic succession for reconstructing sea level is an uninterrupted, highly organic marsh deposit that formed close to high tide level, where accumulation is most tightly controlled by sea-level rise (Allen, 1995, 2000; Gehrels et al., 2005). Hence, we firstly consider such 'stable' stratigraphies using the most organic materials in our compression database. Our modelled stable stratigraphies are uniform, displaying no change in lithology other than minor intra-stratum (stochastic) variability.

A number of recent palaeoecological and numerical studies suggest that the recent sea-level acceleration has resulted in 'transgressive' sediment successions caused by the inability of salt marshes to keep pace with sea level (Donnelly and Bertness, 2001; Kirwan and Temmerman, 2009). In this case, lowering of the salt marsh surface relative to sea level results in an increased minerogenic component and an increased sedimentation rate (Allen, 2000; Pethick, 1981). In our modelled transgressive successions, the stratigraphy records a deepening of water depths; higher marsh deposits are overlain by sediments that accumulate lower in the intertidal frame.

We also consider 'regressive' (shallowing-upward) stratigraphies, which are typical of South Pacific areas where mid- to late-Holocene sea levels were largely stable (Gehrels et al., 2008, 2012). Our modelled regressive stratigraphies record the upwards replacement of tidal flat by salt marsh deposits.

Table 1 provides a summary of modelled stratigraphic successions.

### 4.2. Sub-model 1: generation of synthetic stratigraphies

Sub-model 1 generates synthetic stratigraphic successions. It uses a repeat-iteration, stochastic (Monte-Carlo) approach to explore the effects of natural environmental variability and regression model error on overall compression trends. We run the model 1000 times for each stratigraphic succession considered (see Supplementary information).

There are two model inputs, which we consider deterministically: thickness of the sediment column ($z$, in m) and palaeomarsh

**Table 1**
Summary of synthetic stratigraphic successions generated and decompacted.

| Succession ID | Succession thickness | Stratum 1 | | | Stratum 2 | | | Stratum 3 | | |
|---|---|---|---|---|---|---|---|---|---|---|
| | | Depth range (m bgl) | Mean loss on ignition (%) | Sedimentation rate (mm $yr^{-1}$) | Depth range (m bgl) | Mean loss on ignition (%) | Sedimentation rate (mm $yr^{-1}$) | Depth range (m bgl) | Mean loss on ignition (%) | Sedimentation rate (mm $yr^{-1}$) |
| **Stable/uniform successions** | | | | | | | | | | |
| Acceleration occurs at 0.25 m bgl. Depth range relevant to sea level reconstruction: 0.0–0.48 m bgl. | | | | | | | | | | |
| 1 | 0.5 m | 0.0–0.5 | 39 | 2 | – | – | – | – | – | – |
| 2 | 1 m | 0.0–1.0 | 39 | 2 | – | – | – | – | – | – |
| 3 | 2 m | 0.0–2.0 | 39 | 2 | – | – | – | – | – | – |
| 4 | 3 m | 0.0–3.0 | 39 | 2 | – | – | – | – | – | – |
| **Transgressive successions** | | | | | | | | | | |
| Acceleration occurs at 0.4 m bgl. Depth range relevant to sea level reconstruction: 0.0–0.64 m bgl. | | | | | | | | | | |
| 5 | 1 m | 0.0–0.36 | 35 | 3.3 | 0.36–0.4 | 37 | 3.3 | 0.4–1.0 | 39 | 2 |
| 6 | 2 m | 0.0–0.36 | 35 | 3.3 | 0.36–0.4 | 37 | 3.3 | 0.4–2.0 | 39 | 2 |
| 7 | 3 m | 0.0–0.36 | 35 | 3.3 | 0.36–0.4 | 37 | 3.3 | 0.4–3.0 | 39 | 2 |
| 8 | 1 m | 0.0–0.36 | 33 | 3.3 | 0.36–0.4 | 35 | 3.3 | 0.4–1.0 | 39 | 2 |
| 9 | 2 m | 0.0–0.36 | 33 | 3.3 | 0.36–0.4 | 35 | 3.3 | 0.4–2.0 | 39 | 2 |
| 10 | 3 m | 0.0–0.36 | 33 | 3.3 | 0.36–0.4 | 35 | 3.3 | 0.4–3.0 | 39 | 2 |
| 11 | 1 m | 0.0–0.36 | 30 | 3.3 | 0.36–0.4 | 33 | 3.3 | 0.4–1.0 | 39 | 2 |
| 12 | 2 m | 0.0–0.36 | 30 | 3.3 | 0.36–0.4 | 33 | 3.3 | 0.4–2.0 | 39 | 2 |
| 13 | 3 m | 0.0–0.36 | 30 | 3.3 | 0.36–0.4 | 33 | 3.3 | 0.4–3.0 | 39 | 2 |
| 14 | 1 m | 0.0–0.36 | 27 | 3.3 | 0.36–0.4 | 32 | 3.3 | 0.4–1.0 | 39 | 2 |
| 15 | 2 m | 0.0–0.36 | 27 | 3.3 | 0.36–0.4 | 32 | 3.3 | 0.4–2.0 | 39 | 2 |
| 16 | 3 m | 0.0–0.36 | 27 | 3.3 | 0.36–0.4 | 32 | 3.3 | 0.4–3.0 | 39 | 2 |
| 17 | 1 m | 0.0–0.36 | 24 | 3.3 | 0.36–0.4 | 30 | 3.3 | 0.4–1.0 | 39 | 2 |
| 18 | 2 m | 0.0–0.36 | 24 | 3.3 | 0.36–0.4 | 30 | 3.3 | 0.4–2.0 | 39 | 2 |
| 19 | 3 m | 0.0–0.36 | 24 | 3.3 | 0.36–0.4 | 30 | 3.3 | 0.4–3.0 | 39 | 2 |
| **Regressive successions** | | | | | | | | | | |
| Acceleration occurs at 0.24 m bgl. Depth range relevant to sea level reconstruction: 0.0–0.5 m bgl. | | | | | | | | | | |
| 20 | 1 m | 0.0–0.4 | 39 | 2 | 0.4–0.5 | 24 | 2.5 | 0.5–1.0 | 15 | Not specified |
| 21 | 2 m | 0.0–0.4 | 39 | 2 | 0.4–0.5 | 24 | 2.5 | 0.5–1.0 | 15 | Not specified |
| 22 | 3 m | 0.0–0.4 | 39 | 2 | 0.4–0.5 | 24 | 2.5 | 0.5–1.0 | 15 | Not specified |

*Notes:* standard deviation of loss on ignition values in each stratum were low ( $<1\%$ loss on ignition), even in single iterations of the model.

surface elevation (PMSE, defined in terms of a contemporary altitude in m OD). Each artificial stratigraphic succession is split into layers of equal thickness (0.02 m, the approximate height of a typical oedometer sample). In each layer, the calculated in situ effective stress state and geotechnical properties are assumed to be constant.

A PMSE for each layer is used to generate LOI values, based on established relationships between elevation and LOI (Brain et al., 2011). PMSEs are selected to generate representative downcore LOI profiles, using Cowpen Marsh as our reference study site (site choice does not affect our conclusions). We use the regression models detailed in Section 3 and illustrated in Fig. 4 to generate values of $e_1$, $C_r$, $C_c$ and specific gravity, $G_s$, in each layer. We convert the specified PMSE to a SWLI value to calculate the yield stress, $\sigma'_y$, for each layer using the regression model in Fig. 4(d). Output values from the regression models are selected randomly from statistical error distributions (see Supplementary information). The standard error of the yield stress regression model results in possible predictions of $\sigma'_y < 3$ kPa. Whilst this may be possible, we have no empirical observations of this (Section 3.6). Given the sensitivity of results to low yield stresses, we specify a minimum yield stress value of 3 kPa in all lithologies to prevent extrapolation beyond observed values.

In situ voids ratio values for each layer are calculated using a conditional regression model, formulated as follows:

$$e = e_1 - C_r(\log_{10}\sigma' - \log_{10}\sigma'_y) \text{ if } \log_{10}\sigma' \le \log_{10}\sigma'_y \tag{1a}$$

$$e = e_1 - C_c(\log_{10}\sigma' - \log_{10}\sigma'_y) \text{ if } \log_{10}\sigma' > \log_{10}\sigma'_y \tag{1b}$$

where $e$ is the voids ratio predicted by the model, $e_1$is a constant (the value of voids ratio at 1 kPa, calculated during changepoint regression analysis) and $\sigma'$ is a value of effective stress at the top of the layer under consideration. The logarithmic functions used to describe voids ratio prevent the use of 0 kPa as the value of effective stress at the depositional surface. Instead, we use a value of 0.01 kPa at the depositional surface, which is the order of magnitude of the minimum effective stresses encountered within our modelling experiments (cf. Brain et al., 2011; Smith, 1985).

Beginning with the uppermost layer at the depositional surface, we calculate bulk density ($\rho_d$, in g cm$^{-3}$) as follows:

$$\rho_d = (G_s + e)/(1+e) \tag{2}$$

This permits calculation of the saturated unit weight ($\gamma_{sat}$, in kN m$^{-3}$) of the layer:

$$\gamma_{sat} = \rho_d g \tag{3}$$

where $g$ is the gravitational constant (9.81 m s$^{-2}$). We then calculate the total stress ($\sigma$) at the base of the layer:

$$\sigma = \gamma_{sat} t \tag{4}$$

where $t$ is the thickness of the layer in metres. Assuming hydrostatic conditions, we calculate pore water pressure, $u$, at the base of the layer:

$$u = \gamma_w d \tag{5}$$

where $\gamma_w$ is the unit weight of water (9.81 kN m$^{-3}$) and $d$ is depth ($m$) at the base of the layer below the ground surface. We assume no near-surface unsaturated or capillary saturated zones. Effective stress $\sigma'$ (kPa) acting at the base of the layer can be calculated from

$$\sigma' = \sigma - u \tag{6}$$

Using this value of effective stress, we can then calculate $e$, $\rho_d$, and $\gamma_{sat}$ in the underlying layer. Hence, the effective stress acting at the base of any layer, $\sigma'_{(n)}$, can therefore be calculated from

$$\sigma'_{(n)} = \sigma'_{(n-1)} + ((\gamma_{sat(n)} t_{(n)}) - u_{(n)}) \tag{7}$$

where $\sigma'_{(n-1)}$ is the effective stress at the base of the overlying layer, $\gamma_{sat(n)}$ is the saturated unit weight of the layer, $t_{(n)}$ is the thickness of the layer and $u_{(n)}$ is the pore water pressure at the base of the layer.

For each model run, the outputs of sub-model 1 are entered into sub-model 2.

### 4.3. Sub-model 2: decompaction routine

Sub-model 2 is deterministic, but is run 1000 times for each stratigraphic succession considered, each time using a different set of model values generated in sub-model 1. The decompaction routine involves sequentially removing layers of sediment, beginning with the uppermost layer, to calculate the effective stress profile prior to the deposition of the layer removed. For each layer removed, this involves subtracting the value of effective stress at the base of the layer removed from the values of effective stress in every underlying layer. This provides a new effective stress profile, from which we use the compression model (Fig. 1; Eqs. (1a) and (1b)) to predict values of voids ratio in each layer under the reduced effective stress conditions. Changes in layer thickness with removal of overlying layers can then be calculated using the following equation:

$$\Delta t_{(n)} = \frac{t_{(n)}(e_p - e_s)}{1+e_s} \tag{8}$$

where $\Delta t_{(n)}$ is the change in thickness of a layer, $t_{(n)}$ is the in situ, compacted thickness of the layer, $e_p$ is the model-derived, decompacted voids ratio of the layer and $e_s$ is the in situ, compacted voids ratio.

For each layer removed, the decompaction procedure is completed in every underlying layer. The individual thickness changes in each layer are then summed to calculate the total post-depositional lowering (PDL, m) experienced by the layer immediately below that which has been removed since deposition of overburden sediments. PDL is the height correction that must be added to the in situ altitude of an individual SLI to return it to its depositional altitude.

### 4.4. Sub-model 3: sea level

To demonstrate the effect of compression on sea-level reconstruction, we use an idealised (non-compacted) sea-level curve that contains a recent acceleration, using the depth profile and layer IDs generated earlier in the modelling procedure (Table 1). We assign a year (AD) and sea level (m MSL) to each layer, specifying a pre-inflection rate of 0.2 mm yr$^{-1}$ (cf. Bindoff et al., 2007) and a post-inflection rate of 1.5 mm yr$^{-1}$. We specify an acceleration in the rate of sea-level rise at 1880, with immediate transition from pre- to post-inflection rates. To determine the effect of compression, we subtract the mean PDL ($m$) value for a specific layer from its corresponding sea-level value ($m$). This subtraction results in a distortion of the sea-level record that represents what would be recorded in situ, prior to application of a decompaction correction.

## 5. Results

### 5.1. Post-depositional lowering profiles

In all uniform (stable) stratigraphies, the form of the depth-PDL plots is similar, characterised by no PDL at the top and base of the sediment column and a mid-column peak (Fig. 5). The magnitude of the mid-column PDL peak becomes greater with increases in initial thickness of the sediment column, rising to a

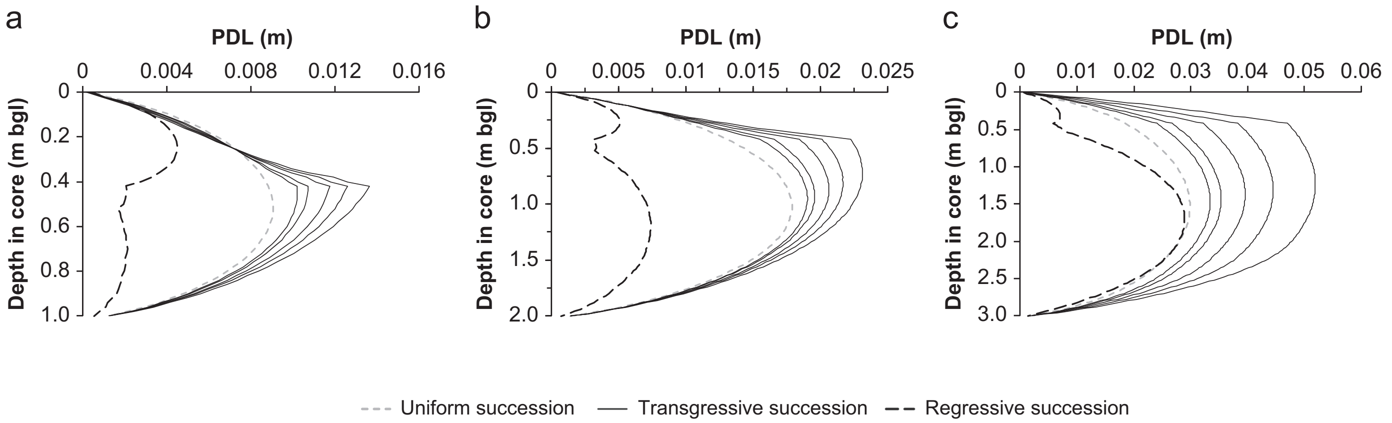


**Fig. 5.** Variations in post-depositional lowering in the modelled cores (1 m, 2 m and 3 m cores only). Note variations in scales of axes. For the transgressive successions, in each graph the mean organic content in the uppermost stratum decreases from left to right: 35%, 33%, 30%, 27% and 24%. For clarity, errors on model predictions are not shown.

maximum of 0.03 m at 1.5 m depth in the 3 m core. The form of the depth-PDL plot is continuously curved, with no obvious inflection point.

For the transgressive stratigraphies considered, we vary the mean LOI in the upper transgressive strata (0.0–0.4 m) between 24% and 35% to demonstrate model sensitivity to this parameter. Where the change in lithology resulting from the transgressive overlap is least pronounced, the overall form of the depth-PDL curve is similar to that observed in uniform stratigraphies with peak PDL at the mid-point of the sediment column (Fig. 5). However, where more marked changes in lithology occur, maximum PDL occurs higher in the core. Again, PDL is greater in deeper stratigraphies (Fig. 5). The maximum value of PDL increases as LOI in the uppermost, transgressive layer decreases (Fig. 5). In addition, as the difference in lithology (LOI) becomes more pronounced at the stratigraphic contact, the form of the PDL plot also changes. Rather than a continuously curved depth-PDL profile, the variation in lithology results in an increasingly pronounced inflection in the PDL plot.

In the regressive stratigraphies modelled (Fig. 5), the magnitude of PDL again increases in thicker stratigraphies, though the PDL is almost always less than in the modelled uniform and transgressive successions. The form of the depth-PDL plots is markedly different. Each stratigraphic layer has a noticeable peak PDL value that is close to, but not at, the centre of each stratum. In the 1 m core, the peak PDL occurs in the uppermost salt marsh layer. As the sediment column thickens and effective stresses exceed the yield stress in the lower mudflat material, the peak PDL shifts to this lower stratum (Fig. 5).

### 5.2. Influence on reconstructed sea level

The changes predicted in Fig. 5 are always less than 0.06 m, but their influence on reconstructed rates of sea-level change is potentially important (Table 2). Fig. 6 shows three examples of the effects of PDL on our synthetic late Holocene/twentieth century sea-level reconstructions. For each we calculated linear rates of sea-level change on the pre- and post-inflection portions of the sea-level curve. For the uniform stratigraphies considered, the total additional contribution of PDL to accelerated sea-level rise ranges from 0.0 to 0.1 mm $yr^{-1}$. The greatest overall additional contributions to acceleration are in the uniform 1 m and 2 m cores. However, in the uniform 3 m core, PDL contributes an additional 0.1 mm $yr^{-1}$ to both pre- and post-inflection rates, though the magnitude of the acceleration remains equal to the true sea-level record.

In the modelled transgressive successions, greater contributions of PDL to accelerated sea-level rise occur in thicker deposits (Table 2; Fig. 7). The magnitude of the compression contribution in all instances increases as the stratigraphic variation in lithology becomes more pronounced. In 1 m thick stratigraphies, the additional contribution does not exceed 0.1 mm $yr^{-1}$, regardless of overburden lithology. In 2 m thick stratigraphies, small changes in lithology also result in an additional 0.1 mm $yr^{-1}$, rising to 0.2 mm $yr^{-1}$ where the transgressive overlap is most marked. In 3 m thick transgressive stratigraphies, our modelling results display a minimum 0.2 mm $yr^{-1}$ additional contribution to the sea-level acceleration. Where changes in organic content (LOI) are most pronounced, the compression contribution rises to 0.4 mm $yr^{-1}$ (Table 2; Fig. 7).

In the modelled regressive stratigraphies, the low PDL values result in the lowest compression contribution to accelerated sea level. No major additional contributions are predicted in such stratigraphies.

## 6. Discussion

### 6.1. Patterns of post-depositional lowering

Greater PDL occurs in thicker stratigraphies due to higher effective stresses generated by increased overburden thicknesses. As effective stresses in lower layers begin to exceed compressive yield stresses, and transitions from low ($C_r$) to high ($C_c$) compressibility behaviour occur, the effect is experienced throughout the sediment column. This is demonstrated by the considerable overall increase in PDL between 2 m and 3 m successions (Fig. 5). The observed mid-column peak in uniform and, to an extent, transgressive successions has previously been reported (Massey et al., 2006; Paul and Barras, 1998; van Asselen et al., 2009) and reflects the optimal combination of loading by overburden sediments and lowering due to compression of underlying material (van Asselen et al., 2009).

The magnitude of PDL in transgressive stratigraphies is considerably greater than in regressive stratigraphies. This results from more compressible sediments being loaded by denser and heavier materials in transgressive successions. In contrast, regressive stratigraphies experience and record the reverse. The greater susceptibility of transgressive stratigraphies to compression relative to their regressive equivalents has been noted previously in both modelling (e.g. Allen, 1999) and field (e.g. Long et al., 2006) studies.

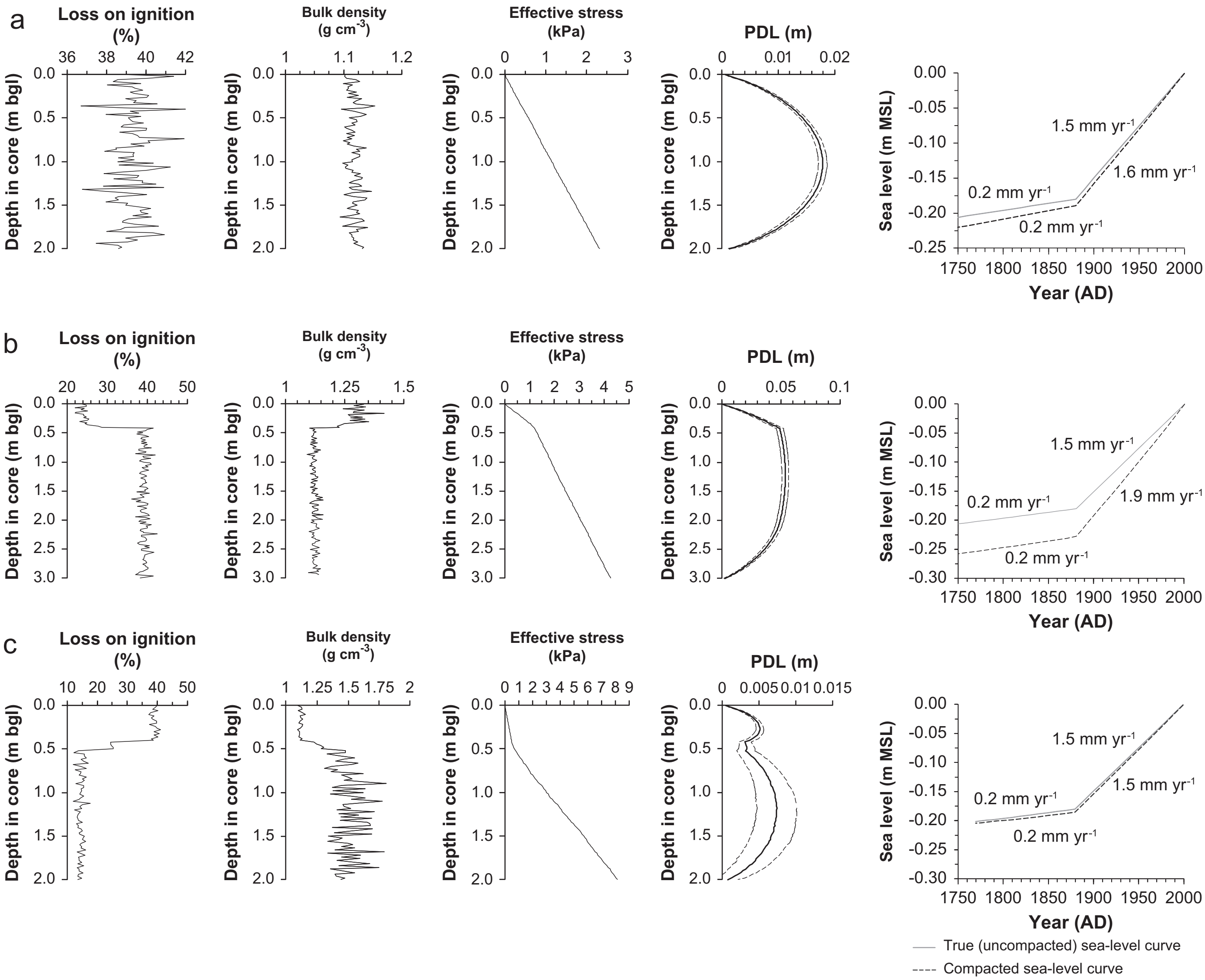


**Fig. 6.** Example model outputs for (a) 2 m uniform succession (Succession ID=3), (b) 3 m transgressive succession (Succession ID=16) and (c) 2 m regressive succession (Succession ID=21). Note differences in scales of axes in each succession. Loss on ignition, bulk density and effective stress profiles are based on single mode iterations to demonstrate similarity with natural cores. Post-depositional lowering (PDL) profiles are based on 1000 model runs and display mean values $\pm 1$ standard deviation for each layer. In the sea level plots, PDL errors are not displayed for clarity, but cluster very closely to the compacted sea level curve.

The form of the PDL profiles varies between transgressive and regressive successions. At transgressive overlaps, changes in compression properties associated with LOI result in marked inflections in PDL. The inflection is more pronounced where a change in lithology, and hence relative density and compressibility between strata, is more marked. In regressive successions, PDL peaks occur within individual stratigraphic units and result from the reduced influence of the compression and lowering of underlying sediments and minimised overburden loading.

### 6.2. *Curved and inflected PDL profiles*

The continuously curved form of the PDL profiles of uniform stratigraphies is significant for our understanding of the causes of the observed acceleration in recent sea-level reconstructed from salt marsh sediments. Were a uniform stratigraphy to accumulate under a regime of steady sea-level rise (no acceleration), the same continuously curved PDL profile would result. Hence, the effect of PDL would result in a continuously curved sea-level plot. In contrast, the nature of the late Holocene/twentieth century acceleration is not gradual with continuous curvature—it is abrupt and distinct (Donnelly et al., 2004; Gehrels et al., 2005, 2006, 2008, 2012; Kemp et al., 2009; Leorri et al., 2008). Hence, in such stratigraphies, compression is unlikely to be the sole cause of the observed acceleration due to the distinct differences in the form of the observed and hypothetical sea-level curves.

In transgressive stratigraphies that demonstrate a large change in organic content at stratigraphic contacts, a sharp inflection in the PDL profile will result in a corresponding inflection of the sea-level curve. We note that transgressive stratigraphies reflect a positive sea-level tendency (or an increase in the proximity of marine conditions; Morrison, 1976), rather than being an indication of an acceleration in the rate of sea-level rise. However, the observation and synchroneity of the inflection at multiple salt marsh sites and in tide gauge records suggest that the existence of a transgressive stratigraphic overlap is itself likely to result from an acceleration in sea-level rise. Hence, whilst compression can contribute to the reconstructed and non-corrected magnitude of acceleration, the acceleration is unlikely to solely be an artefact of local-scale compression processes.

**Table 2**
Summary of the additional contribution of compression/PDL to the magnitude of late Holocene/twentieth century accelerated sea – level rise. Additional compression contributions calculations are based on a modelled 'true' acceleration of 1.3 mm yr – 1.

| Succession ID | Succession thickness (m) | Mean loss on ignition (%) in stratum 1 | Pre-inflection rate (mm $yr^{-1}$) | Post-inflection rate (mm $yr^{-1}$) | Difference (mm $yr^{-1}$) | Additional compression contribution (mm $yr^{-1}$) |
|---|---|---|---|---|---|---|
| **Stable/uniform successions** | | | | | | |
| 1 | 0.5 | 39 | 0.2 | 1.5 | 1.3 | 0.0 |
| 2 | 1 | 39 | 0.2 | 1.6 | 1.4 | 0.1 |
| 3 | 2 | 39 | 0.2 | 1.6 | 1.4 | 0.1 |
| 4 | 3 | 39 | 0.3 | 1.6 | 1.3 | 0.0 |
| **Transgressive successions** | | | | | | |
| 5 | 1 | 35 | 0.2 | 1.6 | 1.4 | 0.1 |
| 6 | 2 | 35 | 0.2 | 1.6 | 1.4 | 0.1 |
| 7 | 3 | 35 | 0.2 | 1.7 | 1.5 | 0.2 |
| 8 | 1 | 33 | 0.2 | 1.6 | 1.4 | 0.1 |
| 9 | 2 | 33 | 0.2 | 1.6 | 1.4 | 0.1 |
| 10 | 3 | 33 | 0.2 | 1.7 | 1.5 | 0.2 |
| 11 | 1 | 30 | 0.2 | 1.6 | 1.4 | 0.1 |
| 12 | 2 | 30 | 0.2 | 1.6 | 1.4 | 0.1 |
| 13 | 3 | 30 | 0.2 | 1.8 | 1.6 | 0.3 |
| 14 | 1 | 27 | 0.2 | 1.6 | 1.4 | 0.1 |
| 15 | 2 | 27 | 0.2 | 1.7 | 1.5 | 0.2 |
| 16 | 3 | 27 | 0.2 | 1.8 | 1.6 | 0.3 |
| 17 | 1 | 24 | 0.2 | 1.6 | 1.4 | 0.1 |
| 18 | 2 | 24 | 0.2 | 1.7 | 1.5 | 0.2 |
| 16 | 3 | 24 | 0.2 | 1.9 | 1.7 | 0.4 |
| **Regressive successions** | | | | | | |
| 20 | 1 | 39 | 0.2 | 1.5 | 1.3 | 0.0 |
| 21 | 2 | 39 | 0.2 | 1.5 | 1.3 | 0.0 |
| 22 | 3 | 39 | 0.2 | 1.5 | 1.3 | 0.0 |

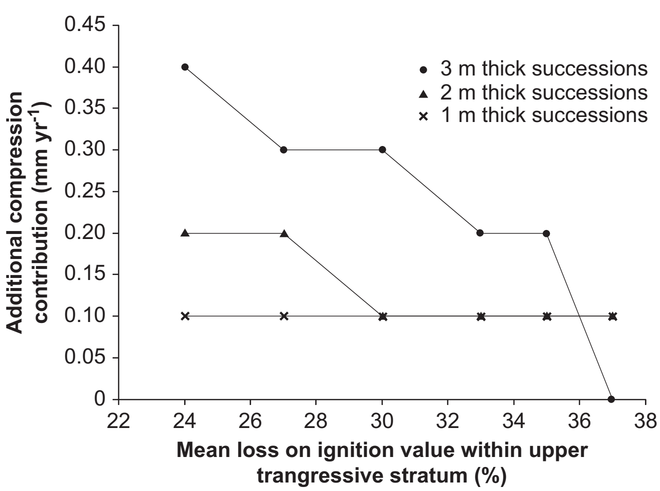


**Fig. 7.** Summary of the modelled total contribution (mm $yr^{-1}$) of compression to the late Holocene/twentieth century acceleration in sea level for 1 m, 2 m and 3 m thick stratigraphic successions.

### 6.3. *Implications for sea-level investigation and interpretation*

The modelled PDL profiles and resultant compacted sea-level curves clearly do not represent all stratigraphies and sea-level histories. Nevertheless, the use of empirical compression datasets, the Monte Carlo approach and the types of stratigraphy modelled allow us to assess the potential contribution of compression to the observed recent acceleration in sea-level rise.

Modelled sea-level reconstructions from transgressive successions that display only subtle variations in lithology are prone to an additional sea-level rise acceleration of 0.1–0.2 mm $yr^{-1}$. This is similar to the average rate at which compaction processes have been estimated to lower SLIs through analysis of basal and intercalated index points (Shennan and Horton, 2002, *ca.* 0.2 mm $yr^{-1}$; Horton and Shennan, 2009, 0.1–0.4 mm $yr^{-1}$). In thicker, more markedly transgressive successions, the additional contribution of compression reached 0.4 mm $yr^{-1}$, which is closer to estimates obtained from thicker ( > 3 m) transgressive stratigraphies of larger UK estuarine systems by Horton and Shennan (2009) (0.6 ± 0.3 mm $yr^{-1}$) and Edwards (2006) (0.7–1.0 mm $yr^{-1}$). However, comparison of our modelled compaction rates with these longer-term stratigraphic studies is not wholly valid, since compaction rates depend on the timeframe considered (decadal, centennial, millennial), sequence depth, lithology and stratigraphy, and the influence of other processes not considered in our model, such as creep and biodegradation, which may be greater over shorter timescales. For example, Törnqvist et al. (2008) reported rates of 5 mm $yr^{-1}$ in the Mississippi Delta in > 5 m sequences over a *ca.* 1500 year period.

The lowest model predictions of the additional contribution of compression to accelerated sea-level rise were observed in shallow uniform and regressive stratigraphies (0.0–0.1 mm $yr^{-1}$). Clearly these are the most suitable environments from which to minimise the potential effects of sediment compaction.

Our modelling shows that the compression contribution to an acceleration in sea-level rise is significant and should not be ignored. Compared to a 'true' increase in the rate of sea-level rise of 1.3 mm $yr^{-1}$, compression-enhanced rates are between 1.4 mm $yr^{-1}$ and 1.7 mm $yr^{-1}$, equal to between 7% and 24% of the observed acceleration. These contributions will be larger where the 'true' sea-level acceleration is smaller. To convert this into real-world impacts, 0.1–0.4 mm $yr^{-1}$ of additional sea-level rise is equivalent to the melting of *ca.* 40–160 $km^3$ $yr^{-1}$ of land-based ice (40–160 Gt $yr^{-1}$ of ice mass loss). The upper values of this range are approaching the equivalent sea-level contribution of ice mass balance loss from the Greenland Ice Sheet in 2006

($250 \pm 40$ Gt yr$^{-1}$ mass loss, equivalent to $0.6 \pm 0.4$ mm yr$^{-1}$; Rignot et al., 2011). Hence, accounting for compression is critical for quantifying the magnitude and causes of sea-level rise and in developing predictive models of future sea-level rise.

Our results show that multiple sea-level records obtained from stratigraphies that vary in thickness and lithological sequence ('stable'/uniform, 'transgressive' or 'regressive') may record large apparent variations in historic sea level, despite an identical 'real' sea-level signal. Constraining the compression contribution is therefore important because global spatial variations in the magnitude of accelerated sea-level rise reconstructed from salt marsh records are similar to those predicted by our modelling experiments (cf. Gehrels et al., 2012; Kemp et al., 2009). Without accounting for compression, such spatial variations could erroneously be attributed to varying cryospheric and/or oceanic forcing, such as the spatial 'fingerprint' of glacier and ice sheet melting resulting from complex interaction of geophysical processes (cf. Mitrovica et al., 2001), or latitudinally-variable steric effects (cf. Wake et al., 2006).

Gehrels et al. (2012) compare salt marsh and tide gauge records of late Holocene/twentieth century sea level and identify a strong latitudinal trend, with faster rates of recent sea-level rise in the southern hemisphere compared with sites further north, leading to the suggestion of a Northern Hemisphere meltwater source for the sea-level acceleration. This spatial variability could, in part, reflect differences in compaction between these sites. The deepest and most strongly transgressive stratigraphies are located on North Atlantic (notably North American) coastlines, whereas the South Pacific salt marshes studied by Gehrels et al. (2008, 2012) comprise regressive stratigraphies, with thin high marsh deposits overlying apparently well-consolidated muds. These Southern Hemisphere records are least prone to compression, yet contain the greatest magnitude of sea-level acceleration. Correcting for compaction of the North Atlantic sites would potentially increase the existing differential between these and the South Pacific sites, amplifying any interpreted Northern Hemisphere melt signal.

### 6.4. Core selection and appraisal of compression

Given the potential problems associated with the use of thick and continuous salt marsh deposits that display variations in lithology throughout the core, the use of compaction-free basal peats provides the best way to limit uncertainty in sea-level reconstructions resulting from sediment compaction (Törnqvist et al., 2004). However, where basal peats are not present, or do not adequately cover the timeframe associated with the late nineteenth to early twentieth century sea-level inflection, our results allow us to make clear recommendations for selection of appropriate stratigraphies. Firstly, thinner sediment columns are preferred, since effective stresses are lower and the cumulative effect of compression in underlying layers is reduced. Secondly, abrupt transgressive contacts should be avoided. Linked to this, LOI and bulk density data for stratigraphies used in reconstruction should be routinely presented and discussed.

Comparison of salt marsh reconstructions with tide gauge records is a powerful tool in assessing the trends and errors in each dataset, allowing the salt marsh method to be validated against a compaction-free record (Gehrels et al., 2012). Whilst salt marsh records may overlap with corresponding tide gauge records in age–altitude plots, the inherent variability in both datasets should be assessed, perhaps with appropriate statistical tests to consider the significance of any differences that may be explained by sediment compression. If no statistically significant differences in trends are observed, sediment compaction is unlikely to have affected the salt marsh record of sea-level change.

Records of dry or bulk density are commonly used to assess the degree of compaction in cores of salt marsh sediment (e.g. Gehrels et al., 2006; Kemp et al., 2009). Constant down-core bulk density is thought to indicate that little or no compaction has occurred. However, our study shows that variability in bulk density within a single stratum resulting from the complex interaction of estuarine, biological and subaerial processes may mask any downcore trend (Fig. 8). This is particularly true in short cores that contain overconsolidated sediments, where the range of effective stresses is small and their absolute values are low. Furthermore, assessing downcore trends in bulk density does not provide any indication of the magnitude and downcore variability in PDL within a core (Fig. 8). This is because PDL depends on the cumulative effect of compression in underlying layers and overburden loading, rather than being solely a function of the effective

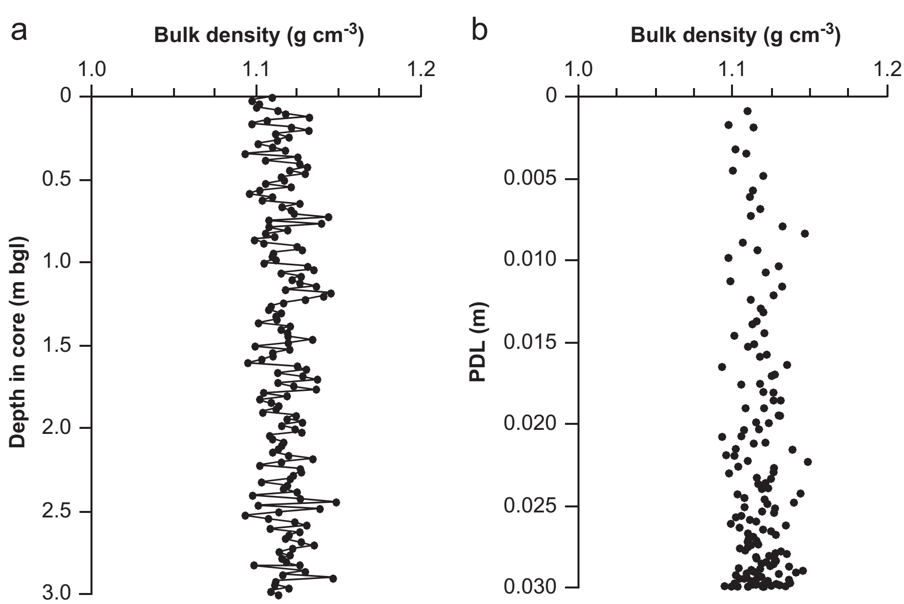


**Fig. 8.** (a) Modelled downcore bulk density profile for a 3 m uniform core (mean loss on ignition=39%). (b) Relationship between modelled bulk density and post-depositional lowering for the same modelled core. These data are for a single model iteration to demonstrate variability in predicted values, as would be observed in real-world cores.

stress state and material properties within a specific layer. Hence, dry or bulk density data provide only a limited insight into potential compaction effects and should not be used to justify 'no compaction' in late Holocene sea-level studies without independent verification.

### 6.5. Future research

Our database is based on primarily minerogenic materials typical of UK and similar NW European salt marshes. Voids ratio and compression index values of the most organic materials in our database are similar to those observed in some amorphous and fibrous freshwater peats (cf. Barden, 1969). However, this similarity is only at the lower end of observed values summarised by Hobbs (1986), who reports maximum values of $C_c$ of 10–12 and $e_i$ of *ca*. 24–26. Should more organic salt marsh deposits demonstrate such compressibility, resultant sea-level curves may be affected to a greater degree. However, compressibility is just one control on PDL. Greater voids ratios (lower bulk densities) result in reduced unit weights and, so, effective stresses within intertidal stratigraphies. Our study shows a strong inverse relationship between bulk density and compressibility. Hence, the loading potential of highly compressible organic materials is likely to be considerably smaller and so the effect of greater compressibility may be offset by the generation of lower effective stresses within a stratigraphic succession. Given the more organic-rich nature of North American salt marshes, and given their importance to current discussions regarding sea-level accelerations, it is important that new contemporary data are collected from these marshes to widen our model application.

A second area of future research is to improve our understanding of the role of processes other than overburden loading that can cause compaction, notably biodegradation and creep. These processes are currently poorly understood in salt marsh deposits, though their importance has been shown to be significant in freshwater peats (Hobbs, 1986) and delta environments (van Asselen et al., 2011).

Finally, the very low stress ($<3$ kPa) compression behaviour of salt marsh deposits represents considerable uncertainty. The mean value of $C_c$:$C_r$ in our database is 7.31 (standard deviation=4.42)—i.e. following exceedance of the compressive yield stress, sediments are $7.31 \pm 4.42$ times more compressible. Hence, if yield stress values $<3$ kPa exist, the high compressibility phase will be active in shorter stratigraphic columns, resulting in greater PDL at shallow depths. However, it is possible that such low values of yield stress may not exist because the zero lateral strain ($K_0$) compressive stress path is anisotropic and requires internal shear deformation (Addis and Jones, 1986). It is therefore possible that the tensile and shear strength of root material (van Eerdt, 1985; Gabet, 1998) is sufficiently high to prevent internal shear at low stresses and so extremely low yield stresses may be unlikely. Nevertheless, constraining low stress behaviour is an important aim for future research.

## 7. Conclusions

We have explored the contribution of sediment compression, a key compaction process, to the magnitude of recent accelerated sea-level rise reconstructed from salt marsh sediments. Our conclusions are as follows:

1. Organic content exerts a key control on the structure, density and compressibility of intertidal sediments. We observe statistically significant relationships between LOI and initial voids ratio and compression indices. These relationships occur regardless of local site conditions, such as variations in geomorphic and hydrographic setting, and ecological and sedimentological character. In contrast, the compressive yield stress displays a more complex relationship with a range of site-specific factors relating to both tidal setting, such as flooding frequency and duration, and eco-sedimentary conditions, such as the presence or absence of surface biomass. Yield stress demonstrates greater dependence on local conditions, but a statistically significant predictive relationship can be defined on a site-by-site basis.
2. Identification of the key controls on compression enables the generation of synthetic stratigraphic successions. We subsequently decompact these successions to assess the effects of sediment column thickness and stratigraphic context ('stable'/uniform 'transgressive' and 'regressive' stratigraphies) on the magnitude of the late Holocene/twentieth century sea-level acceleration reconstructed from salt marsh sediments. Our results show that compression can contribute to, but is unlikely to be the sole cause of, the observed sea-level acceleration.
3. Based on our results, errors associated with compression can largely be avoided by selecting short ($<1$ m) uniform successions, or from thin salt marsh deposits that overlie low compressibility tidal flat deposits. Thicker (2–3 m) sediment columns that display a pronounced 'transgressive' overlap at the time of the acceleration can contribute an additional component of 0.1–0.4 mm $yr^{-1}$of sea-level rise.
4. To draw firm conclusions about the magnitude of the effect of compaction in highly organic sediments ($>50\%$ LOI) requires collection of new data from organic-rich environments and improved understanding of other compaction processes, including creep and biodegradation.

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## Appendix A. Supporting information

Supplementary data associated with this article can be found in the online version at http://dx.doi.org/10.1016/j.epsl.2012.06.045.

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