1. Introduction

Reconstructions of past sea-level changes from the geological record provide important information that can be compared with other proxy climate records as well as output from computer-based models to better understand the past evolution of sea level and its relationship to other components of the climate system. This understanding plays an important role in our ability to project future changes in sea level through empirical scaling relationships (Kemp et al., 2011) and the application of process-based models (Slangen et al., 2012) which are calibrated and tested against palaeo datasets.

Sea-level changes following the last glacial maximum were dominated by the large transfer of water from land ice to the oceans associated with Termination 1. Observational records demonstrate that these sea-level changes had large spatial variability (order 100 m) due to the resulting solid Earth deformation and changes in the Earth’s geopotential associated with the process known as glacial isostatic adjustment (GIA). Accurate interpretation of the sea-level observations therefore requires the application of a GIA model (Peltier and Andrews, 1976; Clark et al., 1978; Nakada and Lambeck, 1989; Mitrovica and Peltier, 1991; Lambeck, 1993). One aspect of this interpretation often involves separating the observed changes into a climate component—that associated directly with mass changes in land ice (the so-called ‘eustatic’ or ‘ice-equivalent’ signal)—and that associated with solid Earth, rotational and gravitational changes. Studies of this type provide constraints on the total volume of land ice lost during Termination 1 as well as the time evolution of this volume loss (Fleming et al., 1998; Lambeck et al., 2002; Milne et al., 2002; Peltier, 2002).

Given that the majority of observational sea-level reconstructions date to Holocene times, a number of studies have focused on constraining ice volume changes during this epoch (Nakada and Lambeck, 1989; Lambeck et al., 2004; Milne et al., 2005). This period is of particular interest as it includes the end of the glacial-interglacial transition during which some large ice sheets disappeared (North American, Eurasian) and the current ice sheets (Greenland, Antarctic) approached a quasi-equilibrium following their primary deglaciation phase. The latter resulted in relatively stable sea levels during the mid-to-late Holocene, a period that represents an ideal reference with which to compare rates of sea-level change in the past century to consider the impact of recent climate warming on sea-level change. The current study adds to the literature on this topic by applying a GIA model to interpret Holocene sea-level observations from the circum-Caribbean region.
The Holocene sea-level history of the circum-Caribbean region has been documented through studies of fossil corals and peat deposits (Toscano and MacIntyre, 2003). The earliest studies from this region (e.g., Currah, 1960; Shepard, 1960) reconstructed sea-level changes using various indicators found in sediments (e.g., shells, wood, peat). These earlier reconstructions are, in general, of relatively low precision. In contrast, the more recent reconstructions based on macrofossils in basal peat deposits provide remarkably accurate and precise constraints on past RSL (Tornqvist et al., 2004). A useful review of the literature on observational records from the circum-Caribbean region can be found in Toscano and MacIntyre (2003), who also presented a sea-level curve for the region comprising existing and unpublished intertidal peat- and coral-based observations, although some of their interpretations were challenged (Blanchon, 2005; Gischler, 2006).

The regional sea-level curve presented in Toscano and MacIntyre (2003) does not account for contributions from either GIA or tectonics. Lambeck et al. (2002) considered a sub-set of data from this region and demonstrated, through the application of a GIA model, that significant spatial differences of several metres in amplitude can exist due to this process. In addition, Milne et al. (2005) demonstrated that the eustatic rise of mid-to-late Holocene RSL in this region is largely a consequence of isostatic deformation due to melting of North American ice. Specifically, the Caribbean and surrounding region has been subsiding during the Holocene due to the collapse of the peripheral bulge associated with the North American glaciation (see Fig. 4 in Milne et al., 2005). Thus, in order to extract an accurate eustatic sea-level signal from the RSL data, these GIA effects (spatial variability and regional subsidence) must be estimated and removed.

The primary aim of this study is to interpret a RSL dataset for the circum-Caribbean region using a GIA model. The main aspects of this interpretation will be to better understand the spatial variability in the observed RSL changes and to remove the influence of GIA from the observations to arrive at an estimate of eustatic changes from 11 cal kyr B.P. to present. We use the entire RSL dataset presented by Toscano and MacIntyre (2003; no data was removed) with the addition of recent data from Belize (Wooller et al., 2004, 2007), the US Gulf coast (Tornqvist et al., 2004), Trinidad (Ramschar, 2004), and the Yucatan Peninsula (Gabriel et al., 2009), as well as new data from Cuba (see Section 2), giving a total of 256 RSL index points. These later sites were added to increase the spatial coverage, an important step given our interest in understanding regional sea level variability. More information about the data can be found in the original publications (e.g., Toscano and MacIntyre, 2003, and references therein).

We provide a brief description of the observations in the next section. In Section 3, we focus on calibrating a GIA model to obtain an optimum fit to the RSL data. We then apply the calibrated model to consider the GIA-induced spatial variability in the region and to estimate and remove the contribution of this process to the data. A broad discussion of the results is given in Section 4 and a summary of our main findings is provided in Section 5.

2. Observations

The geographic distribution of the RSL index points considered is shown in Fig. 1. Relative sea level index points describe a former sea level position as a function of time and consist of two components: an X value (representing an age estimate) and a Y value (representing a former relative sea level position; van de Plassche, 1982; Shennan, 1987; Engelhart and Horton, 2012). All X and Y values also come with associated errors, the calculation of which is described later in this section. In order to present the data as a series of RSL curves, we define 12 localities by averaging the position of each clustered set of index points in a given area. These averaged locations are indicated by the black dots in Fig. 1. Note that some of the index points were not used in this averaging procedure (specifically, those off the west coast of Florida and from Antigua) as they are located too distant from adjacent sites. The data from each of the 12 localities are shown in Fig. 2.

As described above, the two most commonly used sea-level indicators from the circum-Caribbean are mangroves and corals, and we use data from each of these to form the core of our dataset. However, we also consider the RSL curve reconstructed using freshwater peat macrofossils from the Mississippi Delta area, as this proxy has produced sea level data of very high precision, and this extends the spatial coverage of the data (Tornqvist et al., 2004). This is the only site in our study where freshwater samples were used to reconstruct sea level. Throughout this paper, dates are presented in calibrated years before present (cal yr B.P.), with 0 BP representing AD 1950.

2.1. Mangrove peat-based reconstructions

Mangroves are plants that grow along sheltered tropical and subtropical coastlines in the inter-tidal zone (Tomlinson, 1986). In the Caribbean, there are three species of mangrove: red mangrove (Rhizophora mangle), black mangrove (Avicennia germinans), and white mangrove (Laguncularia racemosa), and several associates. The environmental factors that control mangrove distribution are complex and include variables such as frequency of tidal inundation, salinity, elevation, and substrate type (Tomlinson, 1986). Thus, while red mangrove is usually found in a zone closest to the water, it can also grow farther inland, either in estuaries or around inland, brackish bodies of water. Mangroves are generally productive plants, and often build peat, which can accumulate at a rate similar to that of sea level rise (Woodroffe, 1990).

The generation of the relative sea level index points from the mangrove samples was undertaken using the following equation (Engelhart and Horton, 2012):

\[
\text{RSL}_i = \text{RWT} - \text{A}_i
\]

where \(\text{A}_i\) represents the altitude of the sample relative to the tidal datum (e.g., MSL) and \(\text{RWT}_i\) refers to the reference water level. Reference water level is a term that refers to the relationship that the indicator (in this case, mangrove peat) has to a contemporaneous tidal datum (Shennan, 1986), and is defined as the mid-point between the minimum and maximum elevations of the indicator in relation to the same datum as the indicator itself (Engelhart and Horton, 2012). While the relationship between mangrove vertical distribution and sea level can be complex, mangrove peat in the Caribbean generally forms between mean tide level (MLT) and high astronomical tide (HAL; personal communication, Horton, 2012). Thus, in reference to Eq. (1), the RWT for each index point was determined by calculating the mid-point of the vertical range that each indicator occupies for each site based on the tidal information available in the publications that the data were derived from. Relative sea level (RSL) was then determined by subtracting that value from the altitude of the sample relative to the same tidal datum.

The calculation of the vertical error associated with each mangrove-based sea level index point is based on the equation (Shennan and Horton, 2002):

\[
\text{E}_i = \left( e^2_1 + e^2_2 + \ldots + e^2_n \right)^{1/2}
\]

where \(e_1 \ldots e_n\) are individual error terms for sample \(E_i\). The individual error terms used include: (1) the indicative range (or vertical range the indicator occupies); (2) the thickness of the sample in the sediment core; (3) the leveling of the sample during fieldwork; (4) the sampling of the sediment from the core; (5) the benchmark error; and (6), the borehole error. Details of these terms can be found in Engelhart and Horton (2012). The specific values of the individual error terms will vary by site and can be difficult to assess, particularly for data from older publications and sources. When this information was available from the published literature it was used; at sites where this information...
was unavailable, common “conservative” values were estimated. These estimates are 0.1 m for sample thickness, 0.5 m for leveling, 0.01 m for core sampling, 0.1 m for the benchmark error, and 0.01 m for the borehole error (as a percent of depth).

The use of mangrove peat as a sea level indicator can be complicated by several factors. One of the most prominent is auto-compaction—the process by which peat at a certain depth will compact itself due to the weight of the overlying peat (McKee et al., 2007). This has the effect of making compacted peat seem to have been deposited “lower” in terms of its vertical position than peat unaffected by this process. To account for this problem, it may be helpful to consider only sea level index points generated on peat deposited just above an uncompactable surface, such as limestone bedrock. The disadvantage with this approach, of course, is that it reduces the dataset size. A second potential problem with mangrove-based radiocarbon dates is bioturbation—the mixing of younger sediment into older deposits through biological activity, in particular crab burrowing (e.g., Behling et al. 2001). Despite this, there has been limited mention of the bioturbation of mangrove peat in the literature, making its effect difficult to assess.

The age of each mangrove-based sea level index point was determined by recalibrating each date from the original publications using Calib version 6.0 (Stuiver and Reimer, 1993) and the IntCal09 dataset (Reimer et al., 2009). The full two-sigma errors (rounded to the nearest 10 years) were used to calculate each date range.

The following paragraphs describe the peat-based data that were used. Additional details about each site and the associated sea level data can be found in the original publications:

2.1.1. Cuba (Peros, 2005; Davidson, 2007)

The data from Cuba have not been previously published and consist of dates from several cores collected on the north coast of the central part of the island (Fig. 3). The mainland of much of north-central Cuba is separated from the ocean by the Archipiélago de Sabana-Camagüey, a coral reef dominated by a shore-parallel network of limestone and mangrove islands. The Bahía de Buena Vista — the inland body of water adjacent to the mainland (which has a maximum depth of approximately 2 m) — and the Old Bahama Channel to the north are connected through numerous channels between the islands. Tidal modelling results indicate that the tidal range (MHHW–MLLW) in the region is approximately 0.3 m (Hill et al., 2011). Cores were sampled from two broad areas:

(1) Cores D1, D3, and D10 were collected from a chenier plain system west of the town of Punta Alegre. The chenier plain consists of several shore-parallel chenier ridges separated by shallow, interconnected lagoons. The vegetation on and adjacent to the chenier ridges is dominated by the succulent Batis maritima and the mangrove A. germinans. Rhizophora is present within the chenier plain although only in water more than ~0.3 m deep. Throughout this chenier plain system, Avicennia and Batis are found growing at sea level rather than at higher elevations.

(2) Two other cores were collected to the northeast of this chenier plain system. Core CF was sampled on a small limestone island in a dense, monospecific stand of R. mangle. Core PP was sampled in a mixed Rhizophora–Avicennia stand on the landward side of the island of Cayo Guillermo.

Geographic coordinates, elevation data, and radiocarbon information are provided in Table 1. The geographic co-ordinates and elevations were derived from topographic maps produced by the Cuban government. The elevations of the core tops, which were used to derive the depths of the samples below mean sea level, were estimated in the field using tape measures. The following summarizes stratigraphic information for each core:

2.1.1.1. D1 (Chenier plain). A 100 cm core was lifted from the west-central portion of the chenier plain from within a stand of Batis. The core consists of two sedimentary units: (1) a mixture of light-grey silt and shell from 70 to 100 cm; and (2) peat with fine roots and occasional shells from 0 to 70 cm. The 14C date of 300–510 cal yr B.P. was generated on a sample of peat from 60 to 61 cm. The pollen at this level is dominated by Batis (~80%), although there is also some Avicennia pollen (~15%), indicating that a mixed Avicennia–Batis stand was present at the site.
Fig. 2. RSL curves for the 12 localities (black circles) shown in Fig. 1. Each frame includes sea-level index points reconstructed using either fossil corals (Acropora palmata; inverted ‘t-shape’ symbols) or macro- and micro-fossils from peat deposits (black or grey crosses; mangroves except for the Mississippi Delta (MD) reconstruction). The coral data are the least precise with an estimated uncertainty of 5 m, compared to a precision of ±1.0 m for the mangrove peat-based reconstructions. Note that the lower limit of the error bar for the coral-based index points is a hard bound (crossing this bound requires the coral to have lived above mean sea level). In contrast, the upper bound is less well defined since this species of coral has been found living at depths greater than 5 m. Note that the width of the horizontal bar indicating the dating uncertainty. The black crosses identify peat-based index points associated with basal sediment overlying basement rock and so are not affected by compaction. Those recovered from intercalated peat deposits (grey crosses) could have been shifted lower due to this process. See Section 2 for more information on the sea-level index points from each locality. Also shown are the model predictions for the two optimal parameter sets: (1) ICE-5G; upper mantle viscosity (UMV) = $5 \times 10^{21}$ Pas; lower mantle viscosity (LMV) = $5 \times 10^{22}$ Pas (dotted line) and (2) EUST3; UMV = $2 \times 10^{21}$ Pas; LMV = $5 \times 10^{22}$ Pas (dashed line). A lithospheric thickness of 120 km was adopted in each case. For comparison, we also show results for the ICE-5G loading history and the VM2 viscosity model with a 90 km thick lithosphere (grey dotted line).
2.1.1.2. D3 (Chenier plain). Core D3 is 155 cm long. It consists of six stratigraphic units: (1) grey–blue clay from 144 to 155 cm; (2) light grey silt from 119 to 144 cm; (3) from 107 to 119 cm, a layer consisting almost entirely of whole and fragmented shells (shell hash); (4) mineral-rich peat, with thin lenses of shell, from 102 to 107 cm; (5) brown peat with abundant roots from 83 to 102 cm; (6) dark grey–brown organic-rich sediment with some roots from 52 to 83 cm; and (7) fibrous brown peat from 0 to 52 cm. The date of 1180–11510 cal yr B.P. was produced on peat at 103–105 cm consisting of Avicennia (~40%), Batis (~38%) and Rhizophora (~17%) pollen, indicative of an Avicennia–Batis stand.

2.1.1.3. D10 (Chenier plain). Core D10 was sampled from within an Avicennia stand near the western end of the chenier plain. The core is 330 cm long and consists of seven sedimentary units: (1) an olive grey clay, with small pebbles, from 280 to 330 cm; (2) a grey silt unit from 172 to 280 cm; (3) a light grey silt, from 137 to 172 cm, consisting largely of fragmented shells (i.e., shell hash); (4) from 84 to 137 cm, a dark grey, shell-rich mud; (5) from 50 to 84 cm, a unit of similar composition and colour to unit 4, although with roots and fewer shells; (6) from 20 to 50 cm, a grey–brown mud; and (7) a dark brown fibrous peat from 0 to 20 cm. Pollen data indicates that the date of 690–960 cal yr B.P. was made on mud at 122–132 cm deposited in an Avicennia–Batis community.

2.1.1.4. PP (Cayo Guillermo). Core PP is 125 cm long and consists entirely of 125 cm of peat. Small fragments of limestone at the base of the core indicate that limestone bedrock was reached. Fossil pollen was not present in any great abundance in the core but macro-remains at 125 cm, the base of the core, where the date of 680–920 cal yr B.P. was produced, are consistent with red mangrove peat.

2.1.1.5. CF (Cayo Contrabando). A 329-cm core was extracted from a stand of Rhizophora. Small fragments of limestone at the base of the core indicate that bedrock was reached in the coring process. The core consists of sand and coral particles from 266 to 329 cm. Overlying the sand is 266 cm of compact, fibrous, reddish-brown peat containing numerous roots. The colour and texture of the peat throughout the length of the core is similar to modern Rhizophora peat in samples collected from the marsh surface elsewhere on Cayo Contrabando. Pollen is present in the surface samples but exists in very low concentrations in the core, suggesting that conditions are not conducive to its preservation. Nevertheless, the peat on which the basal date of 2350–2710 cal yr B.P. was produced (253–260 cm) appears to have accumulated in a Rhizophora-dominated community.

Additional details on the sites, core stratigraphy, and pollen results can be found in Peros (2005) and Davidson (2007).

### Table 1

<table>
<thead>
<tr>
<th>Sample ID</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Material</th>
<th>$^{14}$C yr B.P.</th>
<th>Cal yr B.P.</th>
<th>Tidal range (m)</th>
<th>Depth of sample in relation to MSL (m)</th>
<th>Reference water level (m)</th>
<th>Relative sea level (m)</th>
<th>Total error (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>D1 60–61</td>
<td>22.3972</td>
<td>–78.8794</td>
<td>Avicennia–Batis peat</td>
<td>360 ± 60</td>
<td>300–510</td>
<td>0.3</td>
<td>–0.55</td>
<td>0.075</td>
<td>–0.625</td>
<td>±0.53</td>
</tr>
<tr>
<td>D3 103–105</td>
<td>22.3900</td>
<td>–78.8643</td>
<td>Avicennia–Batis peat</td>
<td>1420 ± 60</td>
<td>1180–1510</td>
<td>0.3</td>
<td>–1.15</td>
<td>0.075</td>
<td>–1.225</td>
<td>±0.53</td>
</tr>
<tr>
<td>D10 122–132</td>
<td>22.3989</td>
<td>–78.8928</td>
<td>Avicennia–Batis mud</td>
<td>913 ± 75</td>
<td>690–960</td>
<td>0.3</td>
<td>–1.17</td>
<td>0.075</td>
<td>–1.245</td>
<td>±0.53</td>
</tr>
<tr>
<td>PP 125</td>
<td>22.5885</td>
<td>–78.6887</td>
<td>Rhizophora peat</td>
<td>865 ± 55</td>
<td>680–920</td>
<td>0.3</td>
<td>–0.94</td>
<td>0.125</td>
<td>–1.015</td>
<td>±0.53</td>
</tr>
<tr>
<td>CF 253–260</td>
<td>22.5679</td>
<td>–78.7366</td>
<td>Rhizophora peat</td>
<td>2435 ± 55</td>
<td>2350–2710</td>
<td>0.3</td>
<td>–2.47</td>
<td>0.125</td>
<td>–2.545</td>
<td>±0.53</td>
</tr>
</tbody>
</table>

Fig. 3. Map of the Cuba study region showing core locations (black triangles). The 25 m contour is also shown.
2.1.2. Tulum (Gabriel et al., 2009)
Mangrove data from a cenote near the east coast of the Yucatan Peninsula. There is no tide at this site but there is a hydraulic gradient of ~1 m from the Caribbean Sea to the level of the water in the cenote, and we assume that this has remained constant through time. The core used in this study was 0.5 m in length, so it is assumed that compaction was negligible.

2.1.3. Trinidad (Ramcharan, 2004)
Mangrove peat samples from two wetlands in Trinidad. The pollen profiles and descriptions of the sediment cores from this paper show changes in both the vegetation (including red mangrove, “trees and shrubs”, and sedge and grass dominated communities) and sediment composition (from sand, to peaty mud, to silt dominated sedimentary units) at various times, suggesting that the relationship between the peat and mean sea level may be complex. We have decided to retain this site because it provides important spatial coverage in the southeast Caribbean, but note that the data need to be interpreted cautiously.

2.1.4. Jamaica (Digerfeldt and Hendry, 1987)
We used the same mangrove dataset as that of Toscano and Macintyre (2003), which consists of 14 samples from two sites in western Jamaica.

2.1.5. Grand Cayman (Woodroffe, 1981)
This dataset consists entirely of basal red mangrove peat dates. In his paper, Woodroffe (1981) argues that the elevational error is ±0.12 m on the basis of his own leveling and surveying.

2.1.6. Florida (Robbin, 1984)
This is the same dataset used in Toscano and Macintyre (2003), and can be considered to consist of basal samples at each site appears to represent the flooding surface. The dates are made on red mangrove peat.

2.1.7. Belize (Shinn et al., 1982; Macintyre et al., 1995, 2004; Wooller et al., 2004, 2007)
The dataset for Belize includes the data provided in Toscano and Macintyre (2003) for this location. However, we have also added mangrove peat dates from two more recent studies: Wooller et al. (2004), consisting of four AMS dates from a 10-m long core, and Wooller et al. (2007), which consists of nine dates from a ~7.5 m long core. The oldest date in the Wooller et al. (2007) study is from just above bedrock and we have considered it to be a basal date.

2.1.8. Gulf of Mexico (Tornqvist et al., 2004; Yu et al., 2012)
At this location, freshwater macrofossils, rather than mangrove peat, were used to identify palaeo sea level positions. The vertical positions of the sea level index points were corrected for subsidence (due to sediment loading of the Mississippi Delta) at a rate of 0.15 mm/yr (Yu et al., 2012).

2.2. Coral-based reconstructions
We used the coral dataset provided in Toscano and Macintyre (2003). We note that, following the publication of this paper, two comments were published that challenged Toscano and Macintyre’s data selection and interpretation (Blanchon, 2005; Gischler, 2006). Despite these challenges, we have decided to use the Toscano and Macintyre dataset as they make a strong case for why their interpretation of the dataset is reliable (Toscano and Macintyre, 2005, 2006). The calibrated 1- or 2-sigma uncertainties were not recalibrated for the coral data in Toscano and Macintyre (2003), so we recalibrated the conventional 14C ages using the latest version of Calib (Stuiver and Reimer, 1993) and the latest version of the marine calibration dataset (Reimer et al., 2009). For the calibration, the 400-year default reservoir correction was used (Stuiver and Reimer, 1993) and no local delta R value was applied. Like the mangrove data, the full two-sigma errors (rounded to the nearest 10 years) were used to calculate each date range. We note that five of the coral samples (from Florida) from the Toscano and Macintyre (2003) dataset were dated using TLMS U-Th dating and are comparable to the calibrated radiocarbon dates of the other samples.

Following Lighty et al. (1982), we assume that the living depth of Acropora palmata was 0–5 m below mean sea level during the Holocene, although we note that in exceptional cases it has been found at depths of 17 m below sea level (Gorneau and Wells, 1967). Thus, for a sample of fossil recovered at 10 m below contemporary mean sea level, we assign a palaeo mean sea level of 5–10 m. Note that the 10 m value represents the case where the coral was living at mean sea level when it died and so represents a strict lower bound. The upper bound is less strict as this species of coral has been found living at depths considerably below 5 m in the contemporary environment (Zimmer et al., 2006).

For the purposes of using the data to infer optimal model parameters, it is convenient to assign a symmetric 2σ height uncertainty on each data point. This height uncertainty is not calculated in the same manner as the vertical error for the mangrove data (see above), but is based solely on the indicative range of the species. Given the optimal 0–5 m below mean sea level range we are considering, a fossil sample recovered at 10 m below modern sea level, for example, would have a palaeo RSL of R = 7.5 ± 2.5 m. We acknowledge that a symmetric 2σ uncertainty does not accurately represent the true nature of the uncertainty range as indicated above (with a ‘hard’ lower bound and ‘soft’ upper bound). However, the asymmetry is considered when making a visual inspection of the data–model misfit (see Fig. 2 and Section 3.2).

3. Modelling
The focus of this section is to determine a set of model parameters that give the optimum model fit to the dataset described above. This model calibration is performed in Section 3.2. In Section 3.3, we apply the calibrated model to examine the spatial and temporal variability of RSL in the circum-Caribbean region due to GIA. We also correct the RSL observations for GIA effects so that the residual signal more accurately reflects other processes, including changes in ice volume (or eustatic sea level) during the Holocene. We begin this section with a general description of the GIA model components and the key model parameters.

3.1. Model description
The model applied in this study has been described in detail elsewhere and so only the information relevant to the following sections is given. The model comprises three main elements: a model that describes the evolution of grounded ice extent (an ice loading model), a model of solid Earth density and rheology to compute the changes in land height, gravity and rotation associated with the ice-ocean mass re-distribution, and a model that computes sea-level change for a specified ice and Earth model.

Two ice loading models are considered in this study: ICE3G (Peltier, 2004) and EUST3 (Bradley, 2011). ICE3G was developed using a variety of observational constraints from both palaeo reconstructions and contemporary geodetic measurements (Peltier, 2004). The EUST3 model is a significantly revised version of the ICE3G model (Tushingham and Peltier, 1991). The main revisions are the inclusion of a glacial phase and calibration of the model to fit far-field sea-level data during the glacial (Bassett et al., 2005) and Holocene (Bradley, 2011) periods. We note that the EUST3 model has been calibrated only to far-field observations. In contrast, the ICE3G model has been calibrated to both near- and far-field observations.
The Earth model is a spherically-symmetric, compressible, Maxwell body. Models of this type have been employed in a number of sea-level modelling studies (e.g. Peltier and Andrews, 1976; Clark et al., 1978; Wu and Peltier, 1983; Nakada and Lambeck, 1989; Mitrovica and Peltier, 1991; Lambeck et al., 1998; Milne et al., 2005). The response of the model Earth is governed by a suite of so-called visco-elastic Love numbers (Peltier, 1974) which depend on the depth-dependent density and rheology structure of the model Earth. We follow previous studies by adopting a seismic model (Dziewonski and Anderson, 1981) to define the density and elastic properties. The viscous structure of the model is a free parameter set that is varied to optimise the data–model fit (see Section 3.2). Given that the viscosity structure of the Earth is relatively poorly known and RSL data have only a limited depth-resolving power (e.g. Kendall and Mitrovica, 2007), we chose only three parameters to define the viscosity model parameter set: lithospheric thickness (LT), upper mantle viscosity (UMV) and lower mantle viscosity (LMV). Such a crude depth parameterisation is common in GIA studies. The lithosphere is the upper shell of the model Earth in which the viscosity is set to a high value so that this region responds as an elastic plate over timescales of 10–100 kyr. The thickness of this outer shell, which comprises the crust and the upper mantle, is the free parameter in this case. The upper mantle extends from the base of the lithosphere to the seismic mantle boundary (approximately 2900 km depth).

The sea-level model is an extension of the original sea-level equation (Farrell and Clark, 1976) to include time-dependent shoreline migration and an accurate treatment of sea-level changes in regions characterised by retreating marine-based ice (see Mitrovica and Milne, 2002 and references therein). We also include the influence of GIA-induced changes in Earth rotation on sea level (Milne and Mitrovica, 1998; Mitrovica et al., 2006). The algorithm employed is described in Kendall et al. (2005).

While the model applied is relatively sophisticated in most respects, it is important to be aware of its primary limitations. With regard to fitting the observations described above, these limitations are due to simplifications within the parameterisation of the model and the omission of processes that may have made a significant contribution to sea-level change in this region. With regard to the former, the primary simplification of the GIA model is the assumption of 1-D (depth dependent) Earth structure. The circum-Caribbean region is tectonically complex (Burke, 1988) and so there is a distinct possibility that lateral variations in Earth properties will significantly contribute to the GIA signal. A second limitation of the Earth model is the assumption of a linear Maxwell rheology; some recent studies have demonstrated that non-linear flow can have a significant impact on GIA predictions (e.g. Van der Wal et al., 2010). With regard to non-GIA processes that may be important in the study region, the most significant are likely tectonics, palaeo tidal changes, and dynamic sea-surface height changes (associated with changes in ocean density structure and the flow driven by these). The possible influence of these factors with respect to the data–model misfits is discussed in Section 4.

3.2. Model calibration

In this section we seek to determine an optimum parameter set (LT, UMV & LMV) that gives the best fit to the RSL observations for each of the two ice models considered. For each ice model, a total of 243 viscosity models were applied, giving a total of 486 model runs. The Earth models sample a wide range of parameter values that likely encompass those that are plausible — LT: 71 to 120 km; UMV: 0.1–5 × 10^21 Pas; LMV: 1–50 × 10^21 Pas. For each model run, RSL was computed for the location (white circles in Fig. 1) and time of each index point and a data–model misfit was calculated via the relationship,

\[ \chi^2 = \frac{1}{N} \sum_{i=1}^{N} \left( \frac{s_i^o - s_i^p}{\sigma_i} \right)^2 \]  

where N is the number of index points, s_i^o and s_i^p represent, respectively, the observed and predicted values of the ith index point and \( \sigma_i \) is the height uncertainty (95% confidence) in the observed RSL value. The \( \chi^2 \) value indicates the overall quality of the data–model fit for a given model run (the greater the value the worse the fit). Fig. 4 shows the \( \chi^2 \) results for both ice models and the range of UMV and LMV values defined above. Note that the \( \chi^2 \) values were relatively insensitive to changes in LT and so the data cannot discriminate between values in the range considered (71–120 km). The results in Fig. 4 are for a LT of 120 km. Given that the circum-Caribbean region is considered to be an intermediate-field region, it is not surprising that the \( \chi^2 \) results for each ice model are similar (i.e. the predictions are not sensitive to regional differences in the space–time histories of each model). The results for both ice models indicate a clear preference for relatively high viscosity values in the upper and lower mantle regions (as shown by the black contour that indicates the 95% \( \chi^2 \) cut-off value). For the two ice models and the range of viscosity parameters considered, the lowest \( \chi^2 \) values are 6.5 (ICE5G) and 7.8 (EUST3). The \( \chi^2 \) minimum for the ICE5G model was achieved with the viscosity parameters: LT = 120 km; UMV = 5 × 10^21 Pas; LMV = 5 × 10^22 Pas. The minimum for the EUST3 model was achieved using: LT = 120 km; UMV = 2 × 10^21 Pas; LMV = 5 × 10^22 Pas. Using an F-test, the difference between these two \( \chi^2 \) values is not significant at the 95% confidence level but is significant at the 90% confidence level. Most of the results shown below are therefore based on the better-fitting ICE5G loading history.

Given that more precise data have a stronger weighting in computing the \( \chi^2 \) values, it is of interest to compare \( \chi^2 \) results for the coral-based and peat-based datasets separately. Given that the latter are considerably more precise, it is likely that these data dominate the patterns shown in Fig. 4. Fig. 5 shows the \( \chi^2 \) results for the
ICEGS model using only the coral-based reconstructions (top) and the peat-based reconstructions (bottom). In comparison of the results in Figs. 4 & 5, it is clear that the peat-based data do, indeed, dominate the $\chi^2$ results for the entire dataset. We note, however, that while the coral-based data produce a weaker constraint on the model parameters, the general pattern is highly correlated with that of the peat-based data and so the constraints from the two datasets are compatible. That is, both show a preference for relatively high values of UMV and LMV. As an extension of this analysis, we also computed $\chi^2$ values for subsets of the peat-based dataset (Supplementary Fig. 1) with an aim to compare results for the Mississippi data (which is the most accurate and precise of all the data considered) relative to the other peat-based data. The results are similar to those in Fig. 5 in the sense that the Mississippi data clearly produce the best resolved constraint on viscosity structure and the constraints provided by each subset of the peat-based data are compatible. Based on the results in Figs. 4, 5 and Supplementary Fig. 1, we conclude that, while the more precise data dominate the results shown in Fig. 4, the viscosity constraints provided by each subset of the data are compatible and so the preference for relatively high viscosity values is robust.

Fig. 2 shows the RSL curves predicted for both ice models and their optimum (calibrated) viscosity model. At the majority of sites the fits are relatively good but there are a small number of localities where neither model provides a high quality fit. Both model curves sit above the data for Trinidad. There are some basal dates from this locality which indicate that compaction is not a likely explanation for the discrepancy; however, as mentioned in Section 2, there is evidence that the relationship between the peat samples and mean sea level at this locality might have been complex. A similar data–model misfit is evident for Cuba but, in this case, the peat layers are not resting on basement rock and so compaction could be an issue (as could be the relatively crude estimates for tidal range and site elevation at this locality). This also seems to be the case for some of the peat-based reconstructions for Florida, Belize and Jamaica. Neither model captures the Mississippi data during the early part of this record.

We note that, of all the sites considered, four (St Croix/Puerto Rico, Martinique, Trinidad and Panama) are located in proximity to a convergent plate boundary (see, for example, Fig. 1 in Molnar and Sykes, 1969). To make a preliminary investigation of the potential bias in our results due to the influence of vertical tectonic motion, we computed the $\chi^2$ results with the data from these four localities removed. The results (not shown) are very similar to those shown in Fig. 4, with the $\chi^2$ minima located at the same UMV and LMV values when all data are considered and so our inference of viscosity structure appears to be insensitive to the possible influence of vertical tectonic motion at locations proximal to convergent plate boundaries. We return to consider possible causes for these primary data–model misfits in Section 4.

Given that the ICE–5G model was obtained assuming a specific viscosity model (VM2; Peltier, 2004) which includes a depth varying viscosity structure in the upper and lower mantle regions, the model output for this ice–Earth model pairing is also shown on Fig. 2 (grey dotted line). While the fit is similar to that for the optimal Earth model (black dotted line) at some sites, it is clearly worse at a number of others (specifically, MD, FL, CU, TU, BE and TR). This is reflected in the high $\chi^2$ value for this model, which exceeds 200 (note that the depth averages of the VM2 model in the upper and lower mantle regions are approximately $5 \times 10^{20}$ Pas and $5 \times 10^{21}$ Pas, respectively). The poor performance of the ICE–5G/VM2 model pairing in this region is not surprising given that this ice–Earth model combination was calibrated to a global dataset that did not include the RSL data considered in this analysis. It also reflects the fact that significant lateral heterogeneity exists in the real Earth which is not captured in the model applied here or in that employed in the development of ICE–5G. It follows that application of the ICE–5G model with the Earth model that optimises the fit to the Caribbean data considered here would likely produce a poorer fit compared to ICE–5G/VM2 in regions from which data were employed in the development of this latter ice–Earth model pair.

### 3.3. Influence of GIA on RSL in the circum-Caribbean region

In this section, we apply the optimum model inferred above (ICEGS and viscosity model with $L_T = 120$ km; UMV = $5 \times 10^{21}$ Pas; LMV = $5 \times 10^{22}$ Pas) to determine the contribution of GIA to sea-level changes in the study region. In particular, we focus on understanding the spatial variability and temporal aspects of the GIA signal. We conclude this section by presenting a regional sea-level curve in which the GIA component has been removed from the observations.

To illustrate the spatial variability in the GIA signal, we show in Fig. 6 predictions of RSL minus the eustatic component at 10 cal kyr B.P. and 5 cal kyr. These results demonstrate that the spatial variability in this region has changed during the Holocene. During the early Holocene a long-wavelength NNW–SSE gradient in RSL is more apparent. To help interpret the results in Fig. 6, we show in Supplementary Fig. 2, the contribution to this signal from the ice-load induced component of the GIA signal (A) and the ocean-load induced component (B). On comparing the 10 cal kyr B.P. results in Fig. 6 to those in Supp. Fig. 2 (A & B), it is evident that both ice and ocean load changes contribute to the long-wavelength early Holocene gradient, with the ice-induced signal playing the dominant role. Other model output (not shown) indicates that the gravitational influence of the ice-load changes on sea-surface height is the main contributor that drives this RSL gradient. This effect, which is only active as the ice is melting, explains why the gradient is significantly reduced from 8 cal kyr B.P. onwards when most of the ice in North America (and in the applied ice model) had melted.
The results in Supp. Fig. 2 (A & B) show that the ice-induced signal contributes only a long-wavelength (largely N–S) RSL gradient across the circum-Caribbean region, whereas the ocean loading accounts for the more complex signal in which gradients are approximately perpendicular to the modern-day shorelines. GIA-induced changes in Earth rotation produce only a small contribution to the spatial variability across the circum-Caribbean region (see, for example, Fig. 5D in Milne et al., 2005).

In Fig. 7 we show some of the component GIA curves to illustrate how they combine to produce the time variability of the total predicted RSL change (shown for each site in Fig. 2 for each ice model and its optimum partnering viscosity model). While results are shown only for the Cuba locality (black dot labelled ‘CU’ in Fig. 1), the results for other sites are qualitatively similar. We chose Cuba since it lies near the centre of the study region. The solid black curve in Fig. 7 gives the total RSL signal and so is the same as the dotted black line in Fig. 2 (for the Cuba locality). The black dashed curve shows the model eustatic curve and the grey solid and dashed lines show the GIA signals that cause the total RSL change to deviate from the eustatic change. The grey solid line shows the contribution from ice, ocean and rotation signals at this site and the grey dashed line shows the spatially uniform sea-surface height change due to the globally integrated effect of GIA on ocean basin volume change (known as syphoning; Mitrovica and Peltier, 1991; Milne et al., 2002; Mitrovica and Milne, 2002). Note that
these two GIA processes contribute signals of opposite sign and so the deviation from eustasy never exceeds ~5 m since 11 cal kyr B.P. at this locality. The syphoning process dominates until ~9 cal kyr B.P., leading to RSL lying above eustatic sea level in the early Holocene (hence the positive values in Fig. 6A). After this time, the regional influence of ice, ocean and rotational effects becomes more important leading to RSL lying below the model eustatic component (hence the negative values in Fig. 6B).

The dotted grey line in Fig. 7 shows the contribution of ice-load changes to the solid grey line. These changes clearly dominated this signal during the majority of the Holocene. As indicated in Supp. Fig. 2A, this signal is long wavelength and so influences the entire region. By ~8 cal kyr B.P., much of this signal is due to the steady collapse of the peripheral bulge associated with North American deglaciation leading to regional land subsidence and thus enhanced sea level rise. This process is the primary reason that RSL in this region has risen steadily during the late Holocene whereas it has been relatively stable or falling in other low latitude regions (e.g. Clark et al., 1978; Mitrovica and Peltier, 1991).

To end this section, we adopt the best-fitting model to predict and remove the GIA signal from the observations. The residuals will provide a more accurate reflection of eustatic sea-level change, as well as model inaccuracy resulting from non-modelled processes (e.g. tectonics) and limitations in model parameterisation (see Section 4). Fig. 8 (top-left frame) shows all of the ‘raw’ index points (as shown in Fig. 2), while the top frames in middle and right column show, respectively, the ‘raw’ coral and peat-based data. The lower frames show the corresponding RSL residuals after the GIA signal has been removed.

When comparing the ‘raw’ and GIA-revised data in Fig. 8, the most apparent difference is the shift to higher RSL values in mid-to-late Holocene times. As described above, this is largely due to removing the influence of peripheral bulge subsidence. Another point of note is that the revised data still exhibit a considerable degree of scatter, suggesting one or more of the following: (1) non-GIA processes have had a significant influence during the Holocene, (2) aspects of the GIA signal are not accurately captured in the model (e.g. the influence of lateral variations in Earth structure), and (3) data uncertainty is underestimated.

Assuming that the revised data (with GIA removed) provide a more accurate estimate of eustatic sea-level change, we compare the results to the eustatic curves of the two models considered in this study (Fig. 8, bottom-centre and bottom-right frames). Even though the coral data are relatively imprecise, they indicate that the eustatic component of the EUST3 model gives values that fall below the ‘hard’ lower bound on some of the older coral dates. That is, there is too much melting in this model after ~8 cal kyr B.P. (the more melting the deeper the curve goes backwards in time). In this respect the ICE5G model is more compatible with the data from this region. The more precise peak-based data give a similar result. The highest quality peak-based data from the Mississippi Delta area indicate that eustatic sea level continues to rise after 4 cal kyr B.P. (during which time there is effectively zero rise in the ICE5G model). A small amount of ice melt following 4 cal kyr B.P. is included in the EUST3 model as well as some other estimates of late Holocene eustasy (e.g. Fleming et al., 1998; Lambeck and Purcell, 2005). Based on the more precise peak-based data, the GIA corrected values indicate a eustatic rise of 3–4 m from ~7 cal kyr B.P. to ~3–2 cal kyr B.P., with no melting (to within data resolution) from this time. We note, however, that these values are largely based on the data from the Mississippi Delta region and there are some GIA-corrected peak-based index points from other localities that are significantly displaced from these data.

4. Discussion

Given that RSL reconstructions are developed using the height relationship of a given marker (e.g. coral, mangrove) to a certain tidal reference height (e.g. mean high water or mean tide level), it follows that changes in tidal amplitudes through time can affect the accuracy and precision of a given RSL index point. A handful of studies have modelled the changes in tidal range during the most recent deglaciation and have shown that the changes can be significant (~1 m) in some locations (e.g. Uehara et al., 2006; Arbic et al., 2008) and so this could be a mechanism to explain some of the data-model discrepancies shown in Fig. 2 and described at the end of Section 3.2. However, a recent modelling study by Hill et al. (2011) indicates that tidal amplitude changes in the circum-Caribbean region are relatively small during the Holocene. The largest changes (~0.5 m), corresponding to an amplification of the M2 component, occurred during the early Holocene (10–7 cal kyr B.P.). Changes of this amplitude (few 10s of cm) are not large enough to explain the primary misfits shown in Fig. 2.

The contribution to RSL from changes in sea-surface height (SSH) due to shifts in ocean and atmosphere circulation and the associated changes in ocean density structure are not accounted for in this analysis. A small number of studies have considered the impact of these so-called dynamic SSH changes due to melting of the Greenland ice sheet (e.g. Yin et al., 2009; Kopp et al., 2010; Stammer et al., 2011) with a focus on the contribution of this process to future sea-level change. The results of these studies indicate that freshening of the North Atlantic through melt water influx (rates of order 1 mm/yr eustatic sea-level rise) can lead to SSH changes of several decimetres in some regions on decadal to century timescales (largely due to associated changes in the Atlantic meridional overturning circulation). However, we note that, the influence of this process on SSH changes in the circum-Caribbean region is somewhat damped giving signals at the cm level. These studies suggest that the contribution of this process to RSL during much of the Holocene, particularly the mid-to-late Holocene when freshwater input was order 1 mm/yr or less have been small (order cm).

Given that the study region includes a number of plate boundaries at which relative plate motion is being accommodated, it is likely that some sites are influenced to some degree by vertical tectonic motion. Although we demonstrated that the viscosity inference is not significantly altered when a subset of sites located close to a convergent plate margin is removed, this does not preclude the existence of significant vertical motion due to tectonics. By examining the height of coral
terrace formed at the last interglacial relative to present sea level, time-average uplift rates on the order of several decimetres per kyr have been inferred (e.g. Fairbanks, 1989), equating to a signal of order 1–10 m in active areas during the Holocene, which is certainly large enough to account for some of the noted misfits. However, the data–model misfit at Trinidad is curious since it tracks the model predictions quite closely but asymptotes to a present sea level value that is a few metres below mean sea level. This suggests that either the leveling of these data was inaccurate or, if tectonic motion has occurred, it was accommodated by a metre-scale abrupt event (as opposed to more steady changes over the Holocene) within the past millennium.

We cannot rule out contributions from limitations in the GIA model and/or errors in the RSL reconstructions to the data–model misfits shown in Fig. 2. The latter are discussed in Section 2. Regarding the former, the two primary sources of error in the GIA model are inaccuracies in the ice model and Earth model. Given that two different ice models were adopted and the largest misfits (Trinidad, Jamaica) are similar for each, limitations in the Earth model are the most likely primary source of error. This is supported by the fact that the region contains several plate boundaries including subduction zones, and so lateral variations in solid Earth properties are large. Quantifying the model errors in predicted RSL associated with lack of lateral Earth structure has been considered in some previous studies (e.g. Wu and Van der Wal, 2003; Spada et al., 2006; Steffen et al., 2006). These studies show that the influence of lateral structure on Holocene RSL is generally around several 10 s of metres in the near field and several metres in the far field (Spada et al., 2006). Therefore, the application of a GIA model that can incorporate realistic 3-D Earth structure is required to more fully explore the limitations of this analysis and arrive at a more accurate interpretation of the RSL observations.

Both ice models considered produce an optimal fit to the observations only when relatively high viscosities are assigned to the upper and lower mantle (see Fig. 4). These relatively high values are generally greater than those inferred in most GIA studies (e.g. Kaufmann and Lambeck, 2000; Mitrovica and Forte, 2004). The RSL data prefer high viscosity values because the predicted RSL has to lie beneath the eustatic component by a few metres in order to fit the more precise (non-corall) data constraints from ~8 cal kyr B.P. onwards (i.e. the period spanned by the majority of the data considered here). As illustrated in Fig. 7, the departure from eustasy is due to the local contribution from ice, ocean and rotational effects (which contribute a negative signal that is dominated by ice-load-induced peripheral bulge subsidence) and the global drop in mean SSH associated with the syphoning process. The combination of these signals must result in a net rise in RSL during the mid- to-late Holocene and this only occurs for a small class of viscosity models with relatively high UMV and LMV values. The fact that the model fit is highly sensitive to the amplitude of these two component signals raises the issue of model accuracy. For example, how well are they modelled when lateral Earth structure is ignored?

One implication of the relatively poor performance of the ICE-5G/VM2 model in the Caribbean region (see Fig. 2) is the interpretation of last glacial maximum sea levels inferred from Barbados corals (e.g. Peltier and Fairbanks, 2006). The focus of this study is Holocene sea-level change and so this record was not considered. However, it is interesting to note that adopting the preferred viscosity model from this analysis results in a RSL value at the last glacial maximum that is ~13 m shallower compared to the VM2 model. Therefore, an estimate of eustatic sea level based on our regionally calibrated viscosity model would be ~13 m greater compared to that based on VM2.

5. Conclusions

We have used sea-level reconstructions from the circum-Caribbean region (Toscano and Macintyre, 2003; Tornqvist et al., 2004), including unpublished data from Cuba, to calibrate the Earth viscosity model component of a GIA model. Two global ice histories were considered in this calibration process: the widely adopted ICE-5G model (Peltier, 2004) and a more recent model (Bradley, 2011; called EUST3). The two primary objectives in carrying out these viscosity model calibrations were: (1) to consider the contribution of GIA to both the temporal and spatial form of the RSL observations and (2) remove the GIA signal to build upon the study of Toscano and Macintyre (2003) and determine a
more accurate (i.e. GIA-corrected) estimate of eustatic changes from this regional dataset.

For both of the adopted global ice histories, the RSL data were fit best by an Earth viscosity model with relatively high UMV and LMV values, respectively: $5 \times 10^{21}$ Pas and $5 \times 10^{22}$ Pas (ICE-5G), and $2 \times 10^{21}$ Pas and $5 \times 10^{22}$ Pas (EU3T) (see Fig. 4). Considering the results for both of the adopted ice history models, the RSL data constrain UMV to be greater than $0.8 \times 10^{21}$ Pas and LMV to be greater than $3 \times 10^{22}$ Pas (95% confidence). The goodness of fit was unaffected at this confidence level for variations in lithospheric thickness between 71 and 120 km and so the data were unable to define an optimal value for this model parameter (the lowest $\chi^2$ value was produced with a 120 km model lithosphere).

The data preference for such high upper and lower mantle viscosity values is due, largely, to the interplay between two competing component signals – syphoning and peripheral bulge subsidence – and the requirement for the total RSL signal to lie beneath the eustatic component throughout most of the mid-to-late Holocene (Fig. 7). Our results indicate that this requirement can only be satisfied by models with relatively high values of UMV and LMV.

RSL predictions at 10 and 5 cal kyr B.P. were generated using our optimal (calibrated) GIA model (Fig. 6). These indicate considerable spatial variability across the region that is comprised of a long-wavelength north–south gradient and a shorter wavelength signal in which the RSL gradients are perpendicular to coastlines. The former, which has an amplitude of up to 4–5 m across the study region, is due to deglaciation of North American ice and is largest between about 10 and 8 cal kyr B.P., when the Laurentide ice sheet was melting rapidly. The latter is due to ocean loading and produces a signal of several metres between some of the data sites considered in this study during the early to mid-Holocene. When combined these effects result in a spatial variability of RSL during the early Holocene of up to 7 m which is compatible with results from a previous study (Lambek et al., 2002) and large enough to introduce significant error when using the entire dataset to produce a single regional RSL curve (Toscano and MacIntyre, 2003).

On removing the GIA contribution from the best-fitting model, the resulting RSL curve is shifted to shallower depths by several metres for the mid-to-late Holocene (Fig. 8). This is due, largely, to the removal of the peripheral bulge and syphoning effects. Thus, if not corrected for GIA, these data would result in an overestimate of land ice melting during the mid-to-late Holocene. The most accurate and precise data, when corrected for GIA, indicate about 3–4 m of land ice melt (eustasy) since ~7 cal kyr B.P. to ~3–2 cal kyr B.P. (Fig. 8), although we note that there is considerable scatter in the data. Indeed, one surprising outcome of this study is the amount of scatter remaining in the data after the optimal GIA signal is removed. This likely reflects GIA model limitations (particularly the assumption of lateral homogeneity in a region that contains several plate boundaries), inaccuracies in the RSL reconstructions (e.g. due to compaction), and the influence of tectonic processes. Our results indicate that the application of a GIA model that can incorporate the influence of lateral as well as vertical Earth structure would be the most logical extension of this study.

Supplementary data to this article can be found online at http://dx.doi.org/10.1016/j.gloplacha.2013.04.014.

References


Peros, M.C., 2005. Middle to Late Holocene Environmental Change on the North Coast of the Gulf of Mexico. (Unpublished Ph.D. dissertation) University of Toronto.


