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Isolation basin records of late Quaternary sea-level change, central mainland British Columbia, Canada

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ABSTRACT

Isolation basin records from the Seymour-Belize Inlet Complex, a remote area of central mainland British Columbia, Canada are used to constrain post-glacial sea-level changes and provide a preliminary basis for testing geophysical model predictions of relative sea-level (RSL) change. Sedimentological and diatom data from three low-lying (<4 m elevation) basins record falling RSLs in late-glacial times and isolation from the sea by $\sim$11,800–11,200 ^{14}C BP. A subsequent RSL rise during the early Holocene ($\sim$8000 ^{14}C BP) breached the 2.13 m sill of the lowest basin (Woods Lake), but the two more elevated basins (sill elevations of $\sim$3.6 m) remained isolated. At $\sim$2400 ^{14}C BP, RSL stood at 1.49 $\pm$ 0.34 m above present MTL. Falling RSLs in the late Holocene led to the final emergence of the Woods Lake basin by 1604 $\pm$ 36 ^{14}C BP. Model predictions generated using the ICE-5G model partnered with a small number of different Earth viscosity models generally show poor agreement with the observational data, indicating that the ice model and/or Earth models considered can be improved upon. The best data-model fits were achieved with relatively low values of upper mantle viscosity (5 $\times$ 10^{19} Pa s), which is consistent with previous modelling results from the region. The RSL data align more closely with observational records from the southeast of the region (eastern Vancouver Island, central Strait of Georgia), than the immediate north (Bella Bella–Bella Coola and Prince Rupert-Kitimat) and areas to the north-west (Queen Charlotte Sound, Hecate Strait), underlining the complexity of the regional response to glacio-isostatic recovery.

1. Introduction

Late Quaternary relative sea-level records from Canada's north Pacific coast show marked spatial and temporal variation, reflecting elaborate regional differences in the crustal response to deglaciation (Clague et al., 1982; James et al., 2000; Clague and James, 2002; Hutchinson et al., 2004a). At the maximum of the Late Wisconsinan (Fraser) glaciation, the region was occupied by the Cordilleran Ice Sheet, which nucleated in the British Columbia Coast Mountains and flowed into the coastal lowlands, covering Vancouver Island and the Queen Charlotte Islands archipelago (Clague, 1989; Clague and James, 2002; Fig. 1). Deglaciation began at around $\sim$14,000 ^{14}C BP and proceeded rapidly, as frontal retreat at the western periphery of the ice sheet exposed the continental shelf and glaciers stabilized at pinning positions at the front of the Coast Mountains (Clague and James, 2002). By $\sim$10,000 ^{14}C BP ice cover in the region was similar to that of today (Fulton, 1971; Clague, 1981). In the first couple of millennia after deglaciation, RSLs along the mainland coast and in south eastern areas fell rapidly as the crust rebounded, yet in the northwest, the development of a glacio-isostatic forebulge resulted in more complex RSL movements (Josenhans et al., 1995; Barrie and Conway, 2002; Hetherington and Barrie, 2004; Hetherington et al., 2004). Ice sheet decay was accompanied by brief local readvances and by stagnation, which produced further diachroneity in patterns of crustal adjustment (Clague and James, 2002). Relative sea-level records from around the region thus reflect significant variations in ice thickness and distribution, but also the interplay of complex tectonic controls that arise from the proximity of the region to the Cascadia subduction zone (Clague and James, 2002; Hetherington and Barrie, 2004).

Radiocarbon dates from deltaic and marine sediments and terrestrial remains have enabled location-specific RSL curves to be developed for many areas (e.g., Mathews et al., 1970; Andrews and Retherford, 1978; Clague et al., 1982; Hutchinson et al., 2004a). The south mainland coast and eastern and southern parts of Vancouver Island are the most intensely studied (Mathews et al., 1970; Clague

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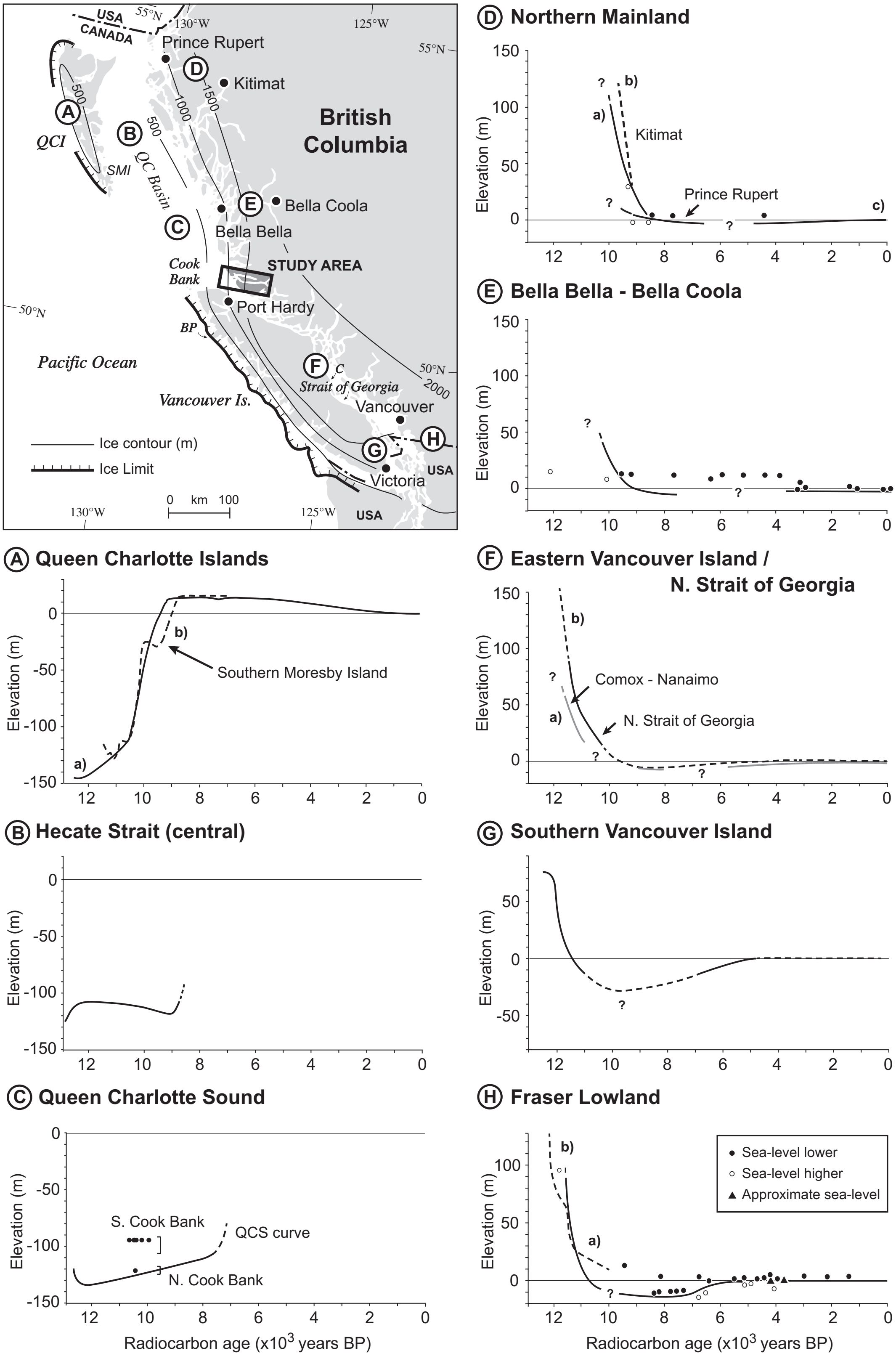

Fig. 1. Post-glacial RSL curves for British Columbia. After: (A) Fedje and Christensen (1999), (Queen Charlotte Islands – a), Hetherington and Barrie (2004) modified from Josenhans et al. (1995), (Southern Morseby Island – b); (B) Hetherington and Barrie (2004); (C) Hetherington and Barrie (2004) with Cook Bank data from Hetherington et al. (2004); (D) Clague et al. (1982) (a, c), Hetherington et al. (2004) (b); (E) Clague et al. (1982); (F) Clague et al. (1982) (Comox-Nanaimo, Eastern Vancouver Island – a), James et al. (2005) (northern Strait of Georgia – b); (G) James et al.'s (2009a) 'preferred' curve (b); (H) Clague et al. (1982) (a), James et al. (2002) (b). All curves are plotted using the ^{14}C time-scale. The map shows the contours of the Cordilleran Ice Sheet at the time of the late Wisconsinan (Fraser) glacial maximum (after Clague, 1983). QCI = Queen Charlotte Islands; QCS = Queen Charlotte Sound; SMI = Southern Morseby Island; QC Basin = Queen Charlotte Basin; BP = Brooks Peninsula; C = Comox. Note: Clague et al. (1982) did not use marine reservoir corrections, thus the RSL fall from the late Pleistocene high-stand position, which in many instances is constrained by marine shell dates, may be up to ∼950 years too old (cf. Hutchinson et al., 2004a, 2004b; James et al., 2009a). This is discussed further in the text.

et al., 1982; James et al., 2000, 2009a; Hutchinson et al., 2004a). Here, the initial rapid uplift of submerged coastal lowlands lasted from approximately 13,000 to 10,000 ^{14}C BP (depending on the locality), and culminated in lower-than-present sea levels between ~9000 and 6000 ^{14}C BP (Clague et al., 1982). The magnitude of the lowstand is not well constrained in many areas, however, and estimates have fluctuated as new information has become available (cf. Mathews et al., 1970; Linden and Schurer, 1988; Mosher and Hewitt, 2004; James et al., 2009a). During the middle and late Holocene, sea levels rose in southern British Columbia gradually to present levels, largely tracking the global eustatic curve (Clague and James, 2002).

Broadly similar patterns are observed along the central and north mainland coast, particularly for the late Pleistocene, although the initial emergence phase appears to have lasted longer, perhaps mirroring later ice retreat (cf. Andrews and Retherford, 1978; Clague et al., 1982; Fig. 1D and E). Differences between outer and inner coast areas (i.e. near fjord heads) have also been noted (Clague et al., 1982).

In marked contrast, RSLs in the Queen Charlotte Islands were well below present between ~13,700 and ~10,000–9500 ^{14}C BP (Hetherington et al., 2004; Fig. 1A). A transgression that began at the end of the Pleistocene peaked about 8500–7500 years ago, when RSLs stood at ~14 m above present (Clague et al., 1982; Josenhans et al., 1995). Emergence then occurred during the last 5000–6000 years (Clague et al., 1982). The notably different RSL histories observed in this region and the adjacent Queen Charlotte Sound area are attributed to thinner ice development at the glacial maximum, earlier deglaciation and to forebulge development (Josenhans et al., 1995; Hetherington and Barrie, 2004), although the spatial extent of this peripheral bulging remain unclear, as no clear equivalent has been recognised to the south of the Queen Charlotte Basin (Hetherington et al., 2004; Fig. 1). This may reflect the fact the forebulge zone was more attenuated in the south, or occurred to the west of Vancouver Island and remains undetected (Hetherington and Barrie, 2004).

While the broad patterns of post-glacial RSL movement around the British Columbia coast are well constrained, wider analysis of regional trends has been hampered by a lack of temporal resolution in the sea-level datasets, particularly for the late Pleistocene (Hutchinson et al., 2004a). The geographical representation of sites is also uneven, and large areas of the coast remain uninvestigated. This is particularly true of remote parts of the central and north mainland coasts, where the isolated and densely forested nature of the terrain have restricted access (Andrews and Retherford, 1978). These limitations have not only hindered detailed studies of isostatic recovery, but have also limited the application of models of crustal displacement (Hetherington et al., 2004), although some work has been undertaken in Vancouver Island (James et al., 2000, 2009a, 2009b).

In this paper we present new RSL data from the Seymour-Belize Inlet Complex (SBIC) on the central mainland coast, an uninhabited fjord complex that lies approximately 40 km to the northwest of Vancouver Island. This area occupies an intermediate position between the Bella Bella–Bella Coola region (~125 km to the north on the mainland coast), and eastern Vancouver Island (>250 km to the southeast), where previously published post-glacial RSL records show several differences (Clague et al., 1982; Fig. 1). Significantly, the area also lies to the immediate south-east of the Queen Charlotte Basin, where, as noted above, inferred forebulge development at the end of the Pleistocene resulted in markedly different RSL movements to those observed on the north mainland coast (Hetherington and Barrie, 2004; Hetherington et al., 2004; Fig. 1B and C). We utilize dated isolation and ingression contacts from the infills of three low-lying coastal basins to construct a preliminary RSL curve for the outer part of the SBIC. The new observational data are used, in turn, to perform a preliminary test of the ICE-5G model of global ice evolution (Peltier, 2004). Specifically, we compare predictions using this model and a small suite of Earth viscosity models to determine if the ice model can capture the observed signal when uncertainties in Earth viscosity structure are considered.

1.1. Study area

The SBIC is characterised by a system of narrow, steep-sided fjords which extend inland by ~70 km, but have a single, narrow (150 m wide) entrance, associated with a shallow (~35 m) sill (Fig. 2). This promotes strong tidal currents at the inlet mouth (the 'Nakwakto Rapids') (Carlson and Hobler, 1976) and restricts access for some vessels. It also reduces circulation, promoting local dysoxia and anoxia of the inlet bottom waters. This has resulted in the preservation of annually resolved marine sediment records for parts of the Holocene (Patterson et al., 2007; Vázquez-Riveiros and Patterson, 2009; Galloway et al., 2010; Babalola et al., 2013). Tidal ranges in the inlets are micromareal; 1.91 m (Mereworth Sound) and 2.0 m (Frederick Sound) (Canadian Hydrographic Service, 2002; Fig. 2). The region is underlain by Mesozoic rocks (mainly granite) and includes mountainous peaks of >2000 m. The steep-sided slopes of the inlets are densely forested and only accessible by boat or air.

The area includes numerous small lake basins within ~1 km of the shore, which range from 0 m to >50 m in elevation. These have the potential to provide important information on post-glacial RSL change. Due to the inaccessibility of the region, these basins have never been investigated, which has hampered an understanding of the deglacial history.

2. Methods

2.1. Site selection

Fieldwork associated with this study was conducted as part of a wider investigation of Holocene climate change in the NE Pacific region, which focused on the collection of marine sediment cores from the SBIC (Patterson et al., 2007; Vázquez-Riveiros et al., 2009). The research reported here was carried out as a parallel investigation, during two week-long cruises to the area in 2002 on the Canadian Hydrographic Survey Vessel, CCGS Vector.

Ten candidate lake basins were selected prior to the field visits from maps and aerial photographs. Low-lying basins (sill heights <10 m) were chosen to elucidate the character of RSL changes during the latter part of the late-glacial and early to mid-Holocene. (For a review of the methodology of acquiring RSL data from isolation basins, see Long et al. (1999, 2011)). Previous research in the central mainland coast in the Bella Bella–Bella Coola area (Andrews and Retherford, 1978; Clague et al., 1982; Fig. 1) has suggested that RSLs remained higher than present until the early Holocene, while data from eastern Vancouver Island and the central Strait of Georgia to the southeast of the region, indicate that slightly higher-than-present RSLs also occurred during the mid-Holocene (Clague et al., 1982; Hutchinson et al., 2004a). Accessibility to the shore was also a critical factor in basin selection, to minimise the distance that equipment had to be transported through the dense temperate rainforest.

Due to logistical constraints, including hazardous tidal conditions, steep terrain and the dense nature of the forest understorey, many of the candidate lakes proved inaccessible. Cores were nevertheless obtained from three basins, Woods Lake, Tiny Lake and Two Frog Lake (informal name) (Fig. 2), with sills of different elevations with respect to present sea level.

2.2. Basin analysis and sediment coring

Depth sounding, using a Knudsen 320BP echo-sounder (28 and 200 kHz), was conducted along two perpendicular transects across the lakes to find the optimal location for coring and to determine infill characteristics (cf. Corner et al., 2001; Balascio et al., 2011). Cores were collected using a Livingstone corer (6 cm diameter) from the deepest or central parts of the lakes, where the sedimentary fill was generally thickest and to avoid potential problems with slumping at the margins.

Sill elevations were surveyed using a Leica 500 differential GPS, supplemented by a total station theodolite in areas where satellite access was poor. Due to the remoteness of the region, there were no permanent survey markers (benchmarks) available to determine absolute datums. As a result, an alternative means of determining sill elevations with respect to present mean tide level (MTL) was required. The coasts around the SBIC are rocky and support barnacles. We used the height measurement of the upper limit of barnacle growth on the modern coast (the 'barnacle line') and related this to MTL (Roe et al., 2009). This method of establishing elevation has been used extensively in Scandinavia (Gray, 1983). In British Columbia, two species of barnacles commonly occur: *Balanus glandula*, which extends to Mean High Water Neap Tide (MHWNT), and *Chthamalus dalli*, which extends up to Mean High Water Spring Tide (MHWST) (Stephenson and Stephenson, 1972; P. Lambert, Royal British Columbia Museum, pers. comm., 2005). Multiple GPS readings were taken of the upper limit of *C. dalli* near the selected basins, which were then averaged to relate the sill elevations to present MTL. Repeated measurements of the barnacle height distribution at several locations around the inlets confirmed that this method of height calculation gave reasonable results when compared with predicted tidal data, supplied by the Canadian Hydrographic Service, for the days of sampling. Total height errors associated with this method are estimated to be no greater than a few centimeters, and the errors are likely to have been constant between sites.

2.3. Sedimentary and microfossil analyses

The sediment cores were logged, X-rayed and sectioned into 1 cm slices for multi-proxy analysis. Samples for diatom analysis were collected at 2–4 cm intervals and at closer intervals across contacts. The sediment organic content was calculated using loss-on-ignition (LOI) analysis, which can assist in identifying isolation and ingression contacts in isolation basins (Anundsen et al., 1994). Samples were air-dried for 24 h at 45 °C and combusted at 550 °C for 4 h (Zong, 1997a). The organic content was calculated as a percentage of the weight loss of the dried samples after combustion (Yu et al., 2003).

For diatom analysis, sediment samples (0.5–2.0 g dry weight) were heated with hydrogen peroxide (30%) until all the organic matter was oxidised and 10% hydrochloric acid was added to remove carbonates, following the method outlined by Battarbee et al. (2001). The diatom suspensions were centrifuged with distilled water to remove chemical residues. The diatoms were mounted in Naphrax and examined using light microscopy ($\times$1000 magnification). A minimum of 500 valves were enumerated per sample. Identifications were made with reference to a range of keys, including Patrick and Reimer (1966, 1975), van der Werff and Huls (1976), Germain (1981), Hartley (1986), Krammer and Lange-Bertalot (1986, 1988, 1991a, 1991b), and Cumming et al. (1995). Diatom classification follows the halobian system of van Dam et al. (1994).

2.4. Radiocarbon dating

The AMS ^{14}C dates were obtained from bulk samples of gyttja around sedimentary contacts and points of diatom-inferred salinity change. Because the dating of marine sediments is influenced by reservoir effects (Kaland et al., 1984), sediment from the freshwater side of isolation contacts was used in most cases. Basal gyttja dates can also be affected by the introduction of old carbon from underlying marine sediments (Hutchinson et al., 2004b). A correction factor of -625 ± 60 years was applied to these dates, following the approach employed by Hutchinson et al. (2004b) and James et al. (2009a) at similar sites in southern British Columbia.

Additional ^{14}C AMS dates were obtained from key points of palynological change in the records as part of a study of the regional vegetational history (Galloway et al., 2007, 2009; Stolze et al., 2007; Table 1). As these were from freshwater sediments (gyttja), they provided a further basis for constraining RSL as 'limiting' dates (cf. Shennan and Horton, 2002); associated RSLs must have been below the basin sills at the time of deposition. The ^{14}C dates were calibrated using IntCal09 (Reimer et al., 2009) and age-depth models were generated using OxCal Version 4.1 (Bronk Ramsey, 2001, 2008; Fig. 3).

2.5. Geophysical model

The model applied in this study has been described in detail elsewhere and so only a general overview is provided here. The model comprises three main components: 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 isostatic response to the ice-ocean mass re-distribution and changes in the rotational potential, and a model that computes sea-level change for a specified ice and Earth model.

We consider the ICE-5G ice loading model in this study (Peltier, 2004). This model was developed using a variety of observational constraints from both palaeo reconstructions and contemporary geodetic measurements (Peltier, 2004). The Earth model is a spherically-symmetric, compressible, Maxwell body. Models of this type have been employed in a number of previous sea-level modelling studies (e.g., Mitrovica and Peltier, 1991; Lambeck et al., 1998; Milne et al., 2005). We follow previous studies by adopting a seismic model (Dziewonski and Anderson, 1981) to define the density and elastic properties. Given that the viscosity structure of the Earth is relatively poorly known, we explore a small number of different viscosity profiles that are defined by three parameters: lithospheric thickness, upper mantle viscosity (UMV) and lower mantle viscosity (LMV). 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 medium over timescales on the order of 100,000 years. For this analysis, we adopted a relatively thin lithosphere of 71 km to be broadly compatible with previous studies that modelled data to the south of the study area (e.g., James et al., 2000). The upper mantle extends from the base of the lithosphere to the seismic velocity increase at 660 km depth; we consider viscosity values of 0.5, 1.0 and 5 $\times$ 10^{20} Pa s for this region. Again, these are relatively low values compared to those inferred from cratonic regions such as eastern Canada and Fennoscandia, but they are broadly compatible with previous studies from the region (e.g., James et al., 2000). The lower mantle extends from this depth to the core-mantle boundary (approximately 2900 km depth); for this region we considered viscosity values of 1, 10 and 50 $\times$ 10^{21} Pa s. This relatively broad range reflects the large uncertainty in viscosity structure within this deep region of the mantle.

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 (2003) and references therein). The

Table 1
Sea-level index points, limiting dates (see text for definition) and age-model inferred data points from the study sites. D = diatomological isolation contact; H = hydrological isolation contact; S = sedimentological isolation contact (for a full definition of these terms and a description of methodological approach used in isolation basin studies see Long et al., 1999, 2011). The ^{14}C dates have been reported previously in the palynological studies of Stolze et al. (2007) (Woods Lake), Galloway et al. (2007) (Two Frog Lake) and Galloway et al. (2009) (Tiny Lake). Rejected dates are italicised. The RSL vertical error calculation includes the sum of all the quantified or estimated height errors, including field levelling (0.1 m), sill elevation (0.25 m), present tide heights and interpretation of indicative meaning (0.2 m) (total error = $\sqrt{(e_1^2 + e_2^2 + \ldots e_n^2)}$) (after Shennan et al., 2005). Note: the lower-most gyttja date from 3.38 m from the Tiny Lake core was not corrected for old carbon effects like the basal gyttja date from Woods Lake because it was 6 cm above the sedimentological isolation contact. HAT = Highest Astronomical Tide; MHWST = Mean High Water of Spring Tides.

a) Isolation basin index points and limiting dates

Basin	Lab. code	Depth (m)	Material dated	^{14}C age yr BP	Corrected ^{14}C age[a]	Calibrated age yr BP (95% CI)	Sill elevation (m) above MTL	Contacts dated	Reference water level	Indicative meaning (m) above MTL	Sea-level tendency	RSL (m) above present MTL	SLI no.
Woods Lake	TO-10780	2.68	Basal gyttja	11,820 ± 90	11,195 ± 108	12,774–13,300	2.13 ± 0.25	H, S	HAT	1.06 ± 0.34	−	1.07 ± 0.34	1
Woods Lake	SUERC-3091	1.43	Gyttja	7176 ± 44	n/a	7878–8154	2.13 ± 0.25	D, H	MHWST-HAT	0.85 ± 0.34	+	1.28 ± 0.34	2
Woods Lake	*TO-10779*	*1.32*	*Gyttja with sand*	*2410 ± 50*	*n/a*	*2344–2702*	*2.13 ± 0.25*	*D, S*	*MHWST-HAT*	*0.85 ± 0.34*	+	*1.28 ± 0.34*	_
Woods Lake	SUERC-4705	1.10	Gyttja with sand	3758 ± 29	n/a	3991–4235	2.13 ± 0.25	D, H, S	MHWST-HAT	0.85 ± 0.34	+	1.28 ± 0.34	3
Woods Lake	*TO-10778*	*0.50*	*Gyttja with sand*	*12,300 ± 90*	*n/a*	*13,935–14,906*	*2.13 ± 0.25*	*D*	*MHWST*	*0.64 ± 0.34*	+	*1.49 ± 0.34*	_
Woods Lake	SUERC-3093	0.19	Gyttja	1604 ± 36	n/a	1404–1565	2.13 ± 0.25	H	HAT	1.06 ± 0.34	−	1.07 ± 0.34	4
Two Frog Lake	*TO-10777*	*4.89*	*Basal gyttja*	*7270 ± 70*	*n/a*	*7953–8285*	*3.59 ± 0.25*	*H*	*HAT*	*1.06 ± 0.34*	−	*2.53 ± 0.34*	_
Two Frog Lake	Beta-185143	4.45	Gyttja	11,040 ± 50	n/a	12,722–13,099	3.59 ± 0.25	n/a	HAT	1.06 ± 0.34	Limiting	n/a	9
Two Frog Lake	Beta-185141	3.10	Gyttja	8620 ± 40	n/a	9529–9675	3.59 ± 0.25	n/a	HAT	1.06 ± 0.34	Limiting	n/a	10
Two Frog Lake	TO-10766	2.93	Gyttja	7550 ± 70	n/a	8189–8509	3.59 ± 0.25	n/a	HAT	1.06 ± 0.34	Limiting	n/a	11
Two Frog Lake	Beta-185142	0.81	Gyttja	2210 ± 40	n/a	2132–2335	3.59 ± 0.25	n/a	HAT	1.06 ± 0.34	Limiting	n/a	12
Tiny Lake	SUERC-3090	3.38	Gyttja	11,763 ± 87	n/a	13,406–13,806	3.58 ± 0.25	H	HAT	1.13 ± 0.34	−?/Limiting	n/a	13
Tiny Lake	TO-12569	1.60	Gyttja	8840 ± 60	n/a	9699–10,169	3.58 ± 0.25	n/a	HAT	1.13 ± 0.34	Limiting	n/a	14
Tiny Lake	TO-12568	1.36	Gyttja	8740 ± 70	n/a	9543–10,121	3.58 ± 0.25	n/a	HAT	1.13 ± 0.34	Limiting	n/a	15
Tiny Lake	BETA-206929	0.88	Gyttja	6860 ± 50	n/a	7592–7819	3.58 ± 0.25	n/a	HAT	1.13 ± 0.34	Limiting	n/a	16

b) Age-model inferred points

	Depth (m)	Estimated ^{14}C age	Estimated calibrated age	Sill elevation (m)	Contacts dated	Reference water level	Indicative meaning	Sea-level tendency	RSL (m) above present MTL	SLI no.
Woods Lake	1.65	8000	8900	2.13 ± 0.25	H	HAT	1.06 ± 0.34	+	1.07 ± 0.34	5
Woods Lake	0.53	2400	2450	2.13 ± 0.25	D	MHWST	0.64 ± 0.34	+	1.49 ± 0.34	6
Woods Lake	0.32	1850	1820	2.13 ± 0.25	D, S	MHWST-HAT	0.85 ± 0.34	−	1.28 ± 0.34	7
Two Frog Lake	4.90	11,800	14,000	3.59 ± 0.25	D, S	MHWST-HAT	0.85 ± 0.34	−	2.74 ± 0.34	8

Sea-level tendency: (+) positive, (−) negative. See van de Plassche (1986) and Shennan et al. (1995) for full definition of terms.
Radiocarbon ages were calibrated using IntCal09 (Reimer et al., 2009) using the 95% confidence limits for the probability option.
Note: Date TO-10780 (Woods Lake, 2.68 m) was incorrectly reported by Stolze et al. (2007) as being based upon plant debris and not gyttja.
[a] Corrections applied: −625 ± 60 yr for bulk organic ages from basal gyttja (following Hutchinson et al., 2004b; James et al., 2009a).

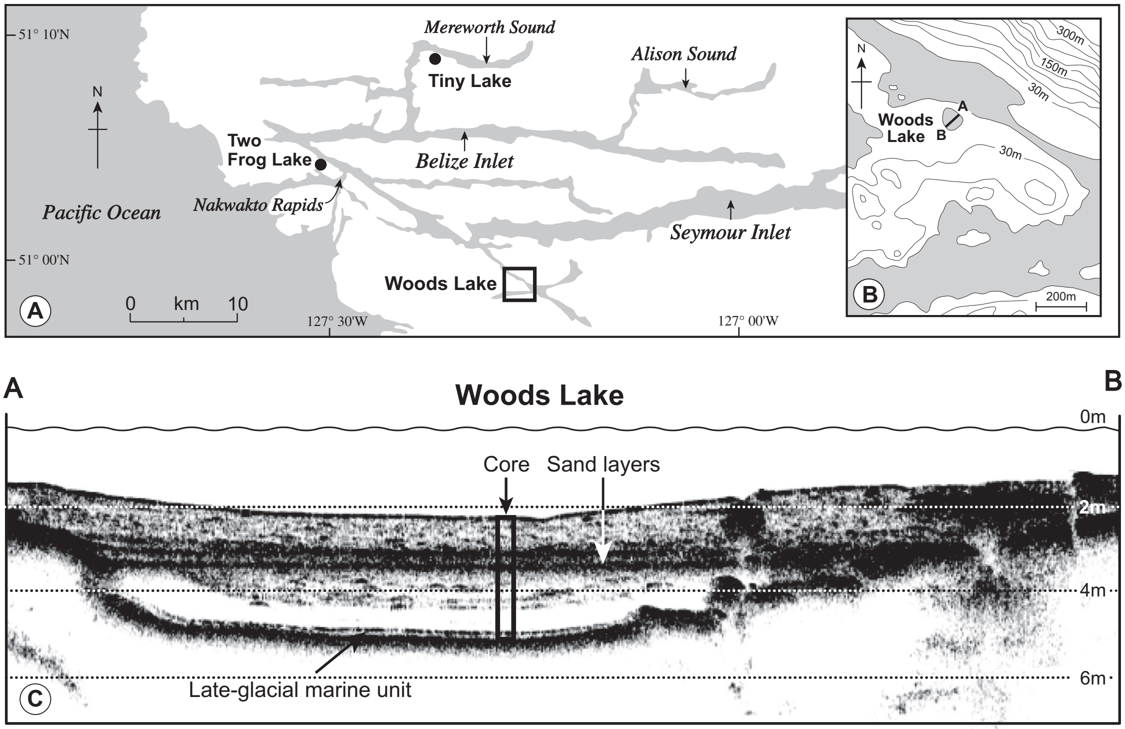

Fig. 2. (A) Map of the Seymour-Belize Inlet Complex (SBIC) showing the three study sites; (B) inset of the Woods Lake basin showing the location of the sub-bottom profile given in C. The sub-bottom profiles for Two Frog and Tiny Lake are given in Appendix B.

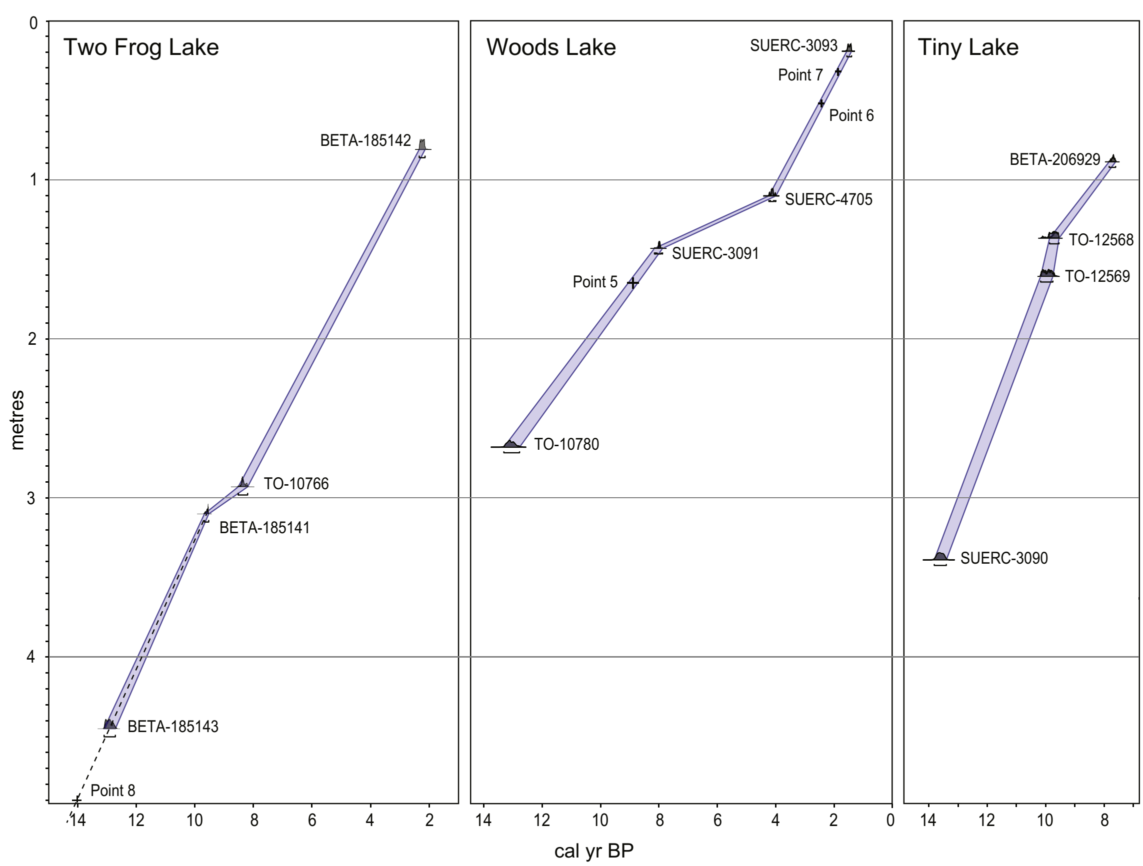

Fig. 3. Age-models for the three lake sediment cores. Details of the sea-level data points and the laboratory codes for the radiocarbon dates are given in Table 1.

algorithm employed is described in Kendall et al. (2005). We also include the influence of GIA-induced changes in Earth rotation on sea level (Milne and Mitrovica, 1998; Mitrovica 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 this study, a key limitation of the GIA model is the assumption of 1-D (depth dependent) Earth structure. Furthermore, the model does not account for tectonic motion.

3. Results and discussion

3.1. Woods Lake

Woods Lake is located 50 m from the southern shore of Seymour Inlet (51°00′15 N, 127°16′07 W) (Fig. 2). It is a small (0.0255 km^2), shallow (c. 2.5 m), closed basin lake with a rock sill that has an average elevation of 2.13 ± 0.25 m above MTL. Six ^{14}C dates were obtained from a 2.82 m sediment core taken from the eastern part of the basin (Stolze et al., 2007; Table 1). Two of these, from brackish water sediments at 0.5 m and 1.32 m, show age reversals and are rejected because of mixing of older (0.5 m) and younger carbon (1.32 m) respectively. A date from similar sediments at 1.10 m plots in reasonable alignment with older dates (Fig. 3), so we cautiously accept this. The date from 1.43 m (7176 ± 44 ^{14}C BP) is further supported by pollen biostratigraphical evidence (it coincides with a regional expansion of Cupresseaceae, which is dated in the Two Frog and Tiny Lake records between ca. 8000 and 7000 ^{14}C BP) (Galloway et al., 2007, 2009; Stolze et al., 2007).

A total of 57 diatom samples were analysed, which yielded diverse assemblages (249 taxa). The diatom diagram (Fig. 4) is subdivided into seven assemblage zones (WL-A to WL-G), based on stratigraphically constrained cluster analysis (Grimm, 1987).

3.1.1. Environmental history

Zone WL – A (2.82–2.76 m): prior to c. 11,195 ^{14}C BP

High frequencies of marine diatoms (>75%) confirm that in late-glacial times the basin was open to the sea. The polyhalobian community is dominated by *Paralia sulcata*, a centric diatom with robust, chain-forming valves (McQuoid and Nordberg, 2003) that tolerates salinities of c. 5–35 ppm (Kjemperud, 1981). Conover (1956) found that in Long Island Sound (USA), *P. sulcata* thrives best in low light conditions in nutrient-enriched waters with temperatures of at least 7 °C. Population numbers of this species can, however, fluctuate when nutrient supply and substrate characteristics, as well as salinities change (Zong, 1997b; Selby et al., 2000; McQuoid and Nordberg, 2003). The rising frequencies of *P. sulcata* throughout the zone may thus partly reflect substrate and water chemistry changes in the basin as it approached isolation.

Zone WL – B (2.76–2.72 m): prior to c. 11,195 ^{14}C BP

An abrupt decrease in polyhalobian and mesohalobian diatoms and a concomitant rise in oligohalobian-indifferent species confirm that this period was associated with significant salinity change, as the basin started to become isolated. It probably still remained connected to the sea during high tides, which permitted low numbers of *P. sulcata* to be carried in or to survive locally. Persistent sea spray may also have introduced marine diatom valves (cf. Schmidt et al., 1990). *Ruppia* pollen occurs in the pollen record, reflecting the local presence of saltmarsh or well-oxygenated, brackish standing waters (Stolze et al., 2007).

The dominant freshwater diatom species include the benthic species *Fragilaria construens* var. *venter*, *Fragilaria exigua* and *Fragilaria pinnata*. The changing proportions probably reflect successional changes during the early formation of the lake (Laing and Smol, 2003). Similar transitions from *P. sulcata* to *Fragilaria*-dominated communities have been recorded in several Scottish (Shennan et al., 1994, 1995; Zong, 1997b; Selby et al., 2000) and Norwegian basins (Kjemperud, 1981) around isolation contacts.

Zone WL – C (2.72–1.68 m): prior to c. 11,195–8200 ^{14}C BP

This zone spans the end of the late-glacial and part of the early Holocene. The 'sedimentological' and 'diatomological' isolation contacts occur at c. 2.68 m (cf. Kjemperud, 1986; Long et al., 1999, 2011) and are associated with a change from grey silts to dark brown gyttja. The LOI values show a corresponding rise (Fig. 4). The presence of a thin (<1 cm), transitional grey-brown silt layer suggests that isolation was relatively abrupt. A corrected basal gyttja ^{14}C date (2.67 m) constrains the age of the contact to 11,195 ± 108 ^{14}C BP.

In the diatom assemblages, *P. sulcata* declines to trace frequencies and oligohalobian-indifferent taxa become dominant. Halophobous diatoms also appear. Small communities of oligohalobian-halophite diatoms (<3–4%) persist to the top of the zone, possibly reflecting salt-water inputs from sea-spray.

Zone WL – D (1.68–1.46 m): c. 8200–7400 ^{14}C BP

Oligohalobian-indifferent species dominate, but decline slightly towards the upper zone boundary. *P. sulcata* re-appears at 1.65 m, rising to 4.6%. Stolze et al. (2007) also reported hystrichospheres (dinoflagellate cysts) in the pollen residues from this part of the core, indicating a renewal of episodic saltwater inputs, probably via over-wash.

Zone WL – E (1.46–0.58 m): c. 7400–2400 ^{14}C BP

Fine sand and scattered small pebbles (<1 cm) occur intermittently in the gyttja associated with this zone (Fig. 4), suggesting that the sea ingressed over the sill periodically, introducing coarse sediment. The sand content first increases near the lower zone boundary and coincides with a decline in LOI values. Interestingly, these slightly coarser facies are recognisable in the sub-bottom profiles, where they appear as a broad reflector that extends evenly across the basin c. 0.5–1.4 m below the lake bed (Fig. 2). The pebbles are also discernable in X-rays (Supplementary Appendix A).

Rising frequencies of mesohalobion diatoms and high frequencies of *Fragilaria brevistriata* and *F. pinnata* indicate that weakly brackish conditions prevailed early in the zone (van der Werff and Huls, 1976). They also confirm that the sill was breached episodically.

An ingression contact, delimited by the first significant reappearance at >5% of *P. sulcata* (Kjemperud, 1981), occurs at 1.30 m. Physical processes, such as tidal mixing might explain the initial increase in *P. sulcata* valves, alongside rising salinities (cf. McQuoid and Hobson, 1998; McQuoid and Nordberg, 2003). Zong (1997b) notes that the ecology of *P. sulcata* is complex, and rising abundances can be associated with both ingression and isolation processes. The ingression may have begun with daily in-washing during the tidal cycle, becoming more pronounced as RSL rose, until *P. sulcata* could occupy niches within the basin.

Zone WL – F (0.58–0.32 m): c. 2400–2000 ^{14}C BP

High frequencies (>45%) of *P. sulcata* suggest that seawater regularly penetrated the basin (Kjemperud, 1981). While this species flourished, other marine and brackish water species were present at only low abundances. This may reflect local nutrient availability or vertical mixing of the water column, which can cause *P. sulcata* to dominate the assemblages in temperate coastal waters (McQuoid and Nordberg, 2003). Interestingly, *P. sulcata* does not attain the same frequencies as it did in zone WL-A, suggesting that access to the sea was more restricted than in late-glacial times. The low end of the salinity range for optimal growth of *P. sulcata* is c. 5–

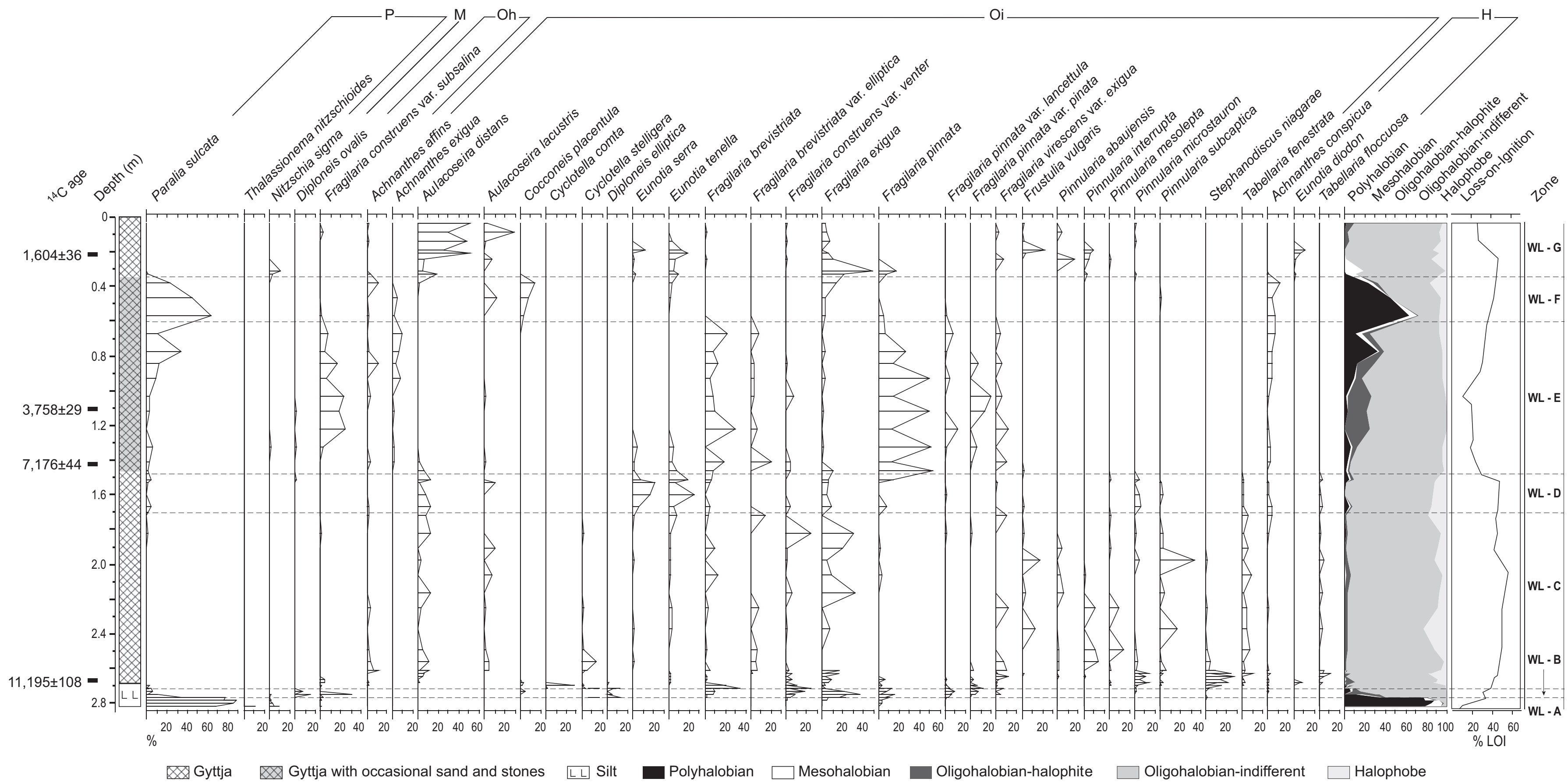

Fig. 4. Diatom percentage diagram for Woods Lake. Only species achieving frequencies >10% are shown. X-rays for the sandy interval of the core between 0.99 and 1.37 m are given in Appendix A; full core logs are given in Doherty (2005). P = polyhalobian, M = mesohalobian; Oh = Oligohalobian-halophite; Oi = Oligohalobian-indifferent; H = Halophobe.

10 ppm; below this, communities are replaced by salt-tolerant freshwater diatoms (Kjemperud, 1986; Zong, 1997b). The observed proportions of *P. sulcata* and *Fragilaria* species probably reflect a gradual reduction in marine influence through the zone and fluctuating salinities. Rising frequencies of *Cocconeis placentula* show that the lake supported aquatic vegetation or fringing saltmarsh communities (Lee et al., 2008).

Zone WL – G (0.32–0.00 m): c. 2000 ^{14}C BP to present

This zone is characterised by significant change as the basin again became isolated. Salt-water inputs ceased early in the zone, as *P. sulcata* declined sharply and freshwater species rose to dominance. This was accompanied by a subtle decline in sand inputs. Fluctuations in *Fragilaria* and *Aulacoseira* species probably reflect changing lake water chemistry, particularly nutrient status and pH, as limnic conditions became re-established. The ^{14}C date based on gyttja from 0.19 m of 1604 ± 36 ^{14}C BP indicates that isolation pre-dated this.

Hystrichospheres disappear from pollen samples in this part of the core, supporting the diatom-based inferences (Stolze et al., 2007). A rise in *Nuphar* pollen towards the top of the zone reflects the expansion of freshwater Nymphaeaceae species (Stolze et al., 2007).

3.2. Two Frog Lake

Two Frog Lake (informal name) is located 100 m from the southern shore of Seymour Inlet (51°06′24 N, 127°32′05 W) (Fig. 2). It is a small (0.08 km^2), shallow (c. 3–5 m), closed-basin lake. The rock sill has an average elevation of 3.59 ± 0.25 m above MTL. A 5.27 m core was collected, which yielded 240 diatom taxa. The diagram is sub-divided into four zones, TFL-A to TFL-D (Fig. 5).

Five ^{14}C dates were obtained (Table 1). A basal gyttja sample (4.89 m) yielded an anomalous age of 7270 ± 70 ^{14}C BP. This is inconsistent with palynological evidence, which suggests that a late-glacial forest community, dominated by *Pinus*, prevailed in the region when the earliest lake sediments were deposited (Galloway et al., 2007). This date is therefore rejected. The age model for the core (Fig. 3) is based on the four remaining dates.

3.2.1. Environmental history

Zone TFL-A (5.25–4.91 m): c. 12,000 ^{14}C BP

The well-developed marine and brackish diatom communities, which include *Coscinodiscus nodulifer*, *Nitzschia lanceolata*, *Thalassiosira baltica*, and *Surirella ovalis*, show that the basin was open to the sea. The presence of the epipsammic species *Cocconeis peltoides*, is consistent with the silts and fine sands that form the associated sediments. A gradual decrease in polyhalobian and mesohalobian diatoms and a rise in oligohalobian-indifferent species (particularly *F. pinnata*), suggest that salt-water penetration became more restricted near the top of the zone.

Zone TFL-B (4.91–2.72 m): c. 12,000–7000 ^{14}C BP

The abrupt replacement of light grey silts by mid-brown gyttja at the base of the zone indicates that the basin was isolated rapidly. This is also reflected in increasing LOI values between 4.90 and 4.80 m (Fig. 5). These changes were accompanied by the sudden disappearance of polyhalobian and mesohalobian diatoms and an increase in oligohalobian taxa. Rising frequencies of halophobous species at 4.70 m, suggest that freshwater conditions quickly became established.

Zone TFL-C (2.72–0.50 m): c. 7000–c. 1400 ^{14}C BP

Quiet, freshwater conditions prevailed. Massive dark brown gyttja forms the associated sediment. Occasionally fine sand was introduced, producing small fluctuations in the LOI values (Fig. 5). The sub-bottom profiles do not record any significant changes in the lithostratigraphy, although a thin, continuous reflector corresponding to c. 1.7–1.8 m core depth may represent a concentration of clastic sediments around this horizon (Supplementary Appendix B). Changes in the halophobous diatom community at this point may also reflect substrate-related changes (Fig. 5).

Zone TFL-D (0.50–0.00 m): c. 1400 ^{14}C BP to present

Freshwater conditions persisted. Low frequencies of polyhalobian diatoms (up to 3–4%) might reflect intermittant inputs from sea-spray rather than episodic over-wash events, which would probably have had a larger impact on the halophobous diatom communities.

3.3. Tiny Lake

Tiny Lake is located 200 m from the southern shore of Mereworth Sound, Belize Inlet (51°11′40 N, 127°22′48 W) (Fig. 2). It is a small (0.48 km^2), relatively deep lake (up to 20 m), with a shallow rock sill, which has an average elevation of 3.58 ± 0.25 m above MTL. The current catchment has one small stream, which flows over the sill on the north-eastern edge of the lake and out to the sea. A 3.56 m sediment core was collected from the south-eastern side of the basin. An age model for this core is based on four AMS ^{14}C dates (Table 1; Fig. 3). The diatom diagram, with three assemblage zones (TL-A to TL-C), is presented in Fig. 6.

3.3.1. Environmental history

Zone TL-A (3.56–3.45 m): c. 12,000 ^{14}C BP

High frequencies of the shallow marine diatom *C. nodulifer*, suggest that the basin had open access to the sea. *Diploneis didyma* is an epipelic species commonly found in brackish mud-flats (Vos and de Wolf, 1993). Its increase above 3.48 m may be a response to a change in substrates, in particular a shift from sandy silt to silty clay aggradation. A rise in the fresh-brackish aerophilous species *Diploneis ovalis* at 3.46 m probably reflects the local development of high saltmarsh environments in and around the basin (Vos and de Wolf, 1993; Gehrels et al., 2001).

Zone TL-B (3.45–1.43 m): c. 12,000–5000 ^{14}C BP

The replacement of silty clays by gyttja and a sharp decline in polyhalobian diatoms at the base of the zone indicate that the sea withdrew rapidly. The ‘diatomological’ and ‘sedimentological’ isolation contacts both occur at c. 3.45–3.44 m. *D. ovalis* still remained present early in the zone, peaking at 3.44 m (2%) and then declining to trace levels above 3.36 m. This suggests that saltmarsh communities persisted locally for a short period after isolation. The observed changes in the *Fragilaria* species early in the zone probably reflect successional changes in the newly formed lake as substrate and nutrient status changed alongside declining salinities.

Scattered fine sand occurs in the gyttja at c. 2.00–2.22 m and 1.43–1.68 m and corresponds to small-scale fluctuations in LOI values (Fig. 6; Appendix B). No evidence of renewed salt-water penetration is recorded in the diatom communities, however.

Zone TL-C (1.43–0.00 m): c. 5000 ^{14}C BP–present

Freshwater conditions prevailed. Small changes in the dominant oligohalobian-indifferent and halophobous communities around the zone boundary may reflect substrate-related changes.

3.4. Relative sea-level reconstruction

3.4.1. Late-glacial

In late-glacial times (prior to c. 11,195 ^{14}C BP) the three basins were fully connected to the sea (Fig. 7). The onset of marine conditions is

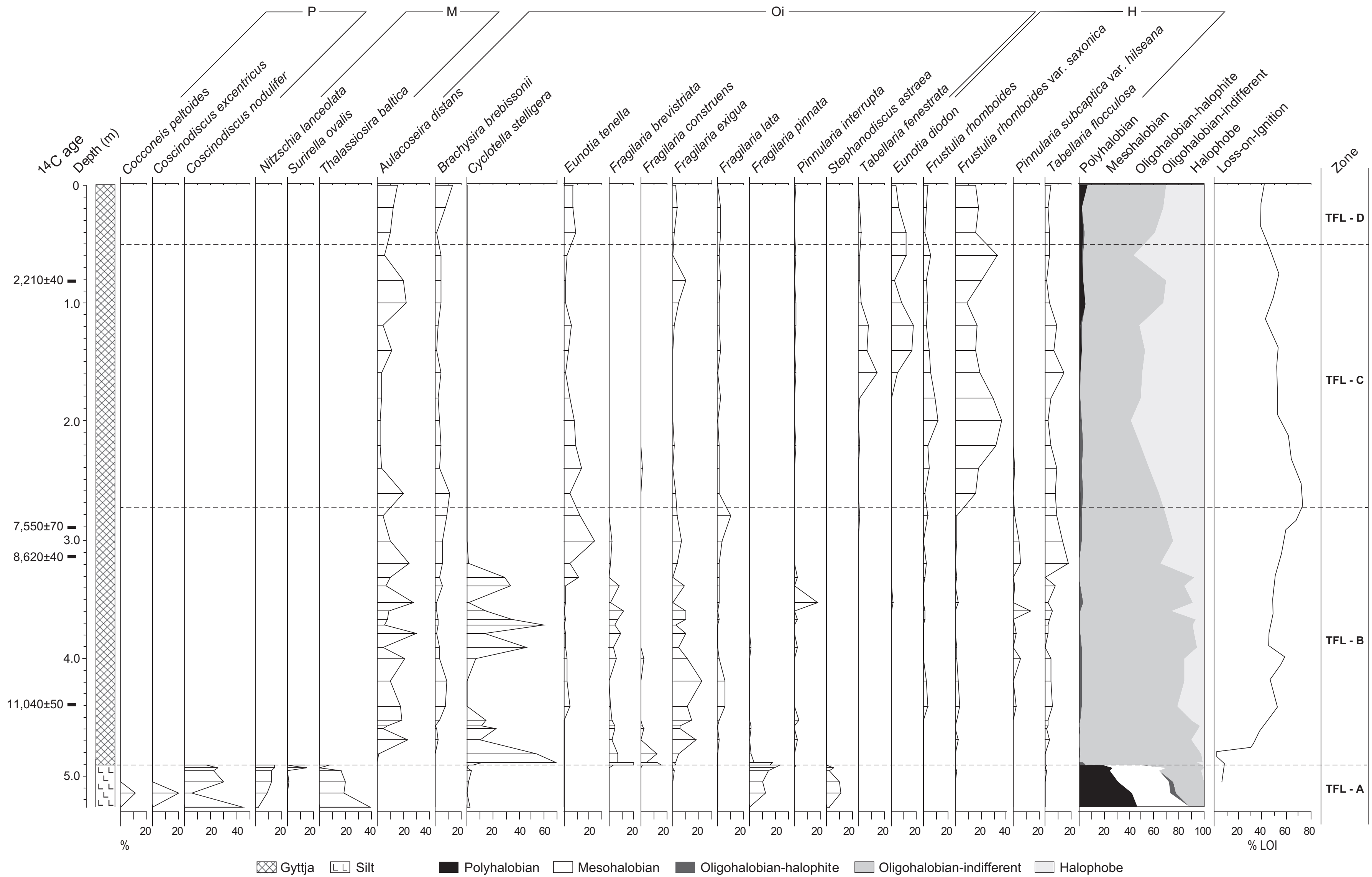

Fig. 5. Diatom percentage diagram for Two Frog Lake. Only species achieving frequencies >10% are shown. Abbreviations for the salinity classes follow Fig. 4.

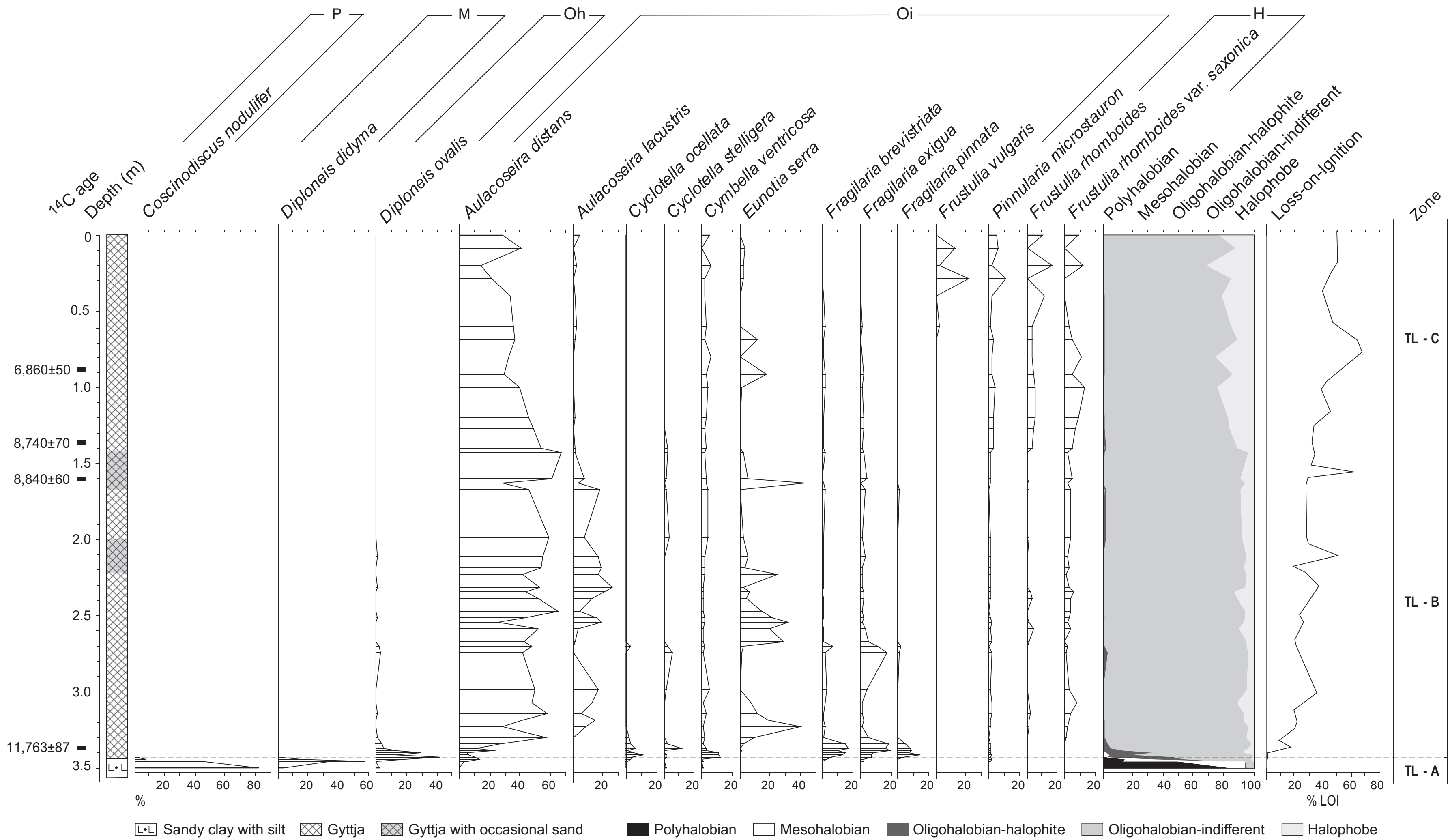

Fig. 6. Diatom percentage diagram for Tiny Lake. Only species achieving frequencies >10% are shown. Abbreviations for salinity classes follow Fig. 4.

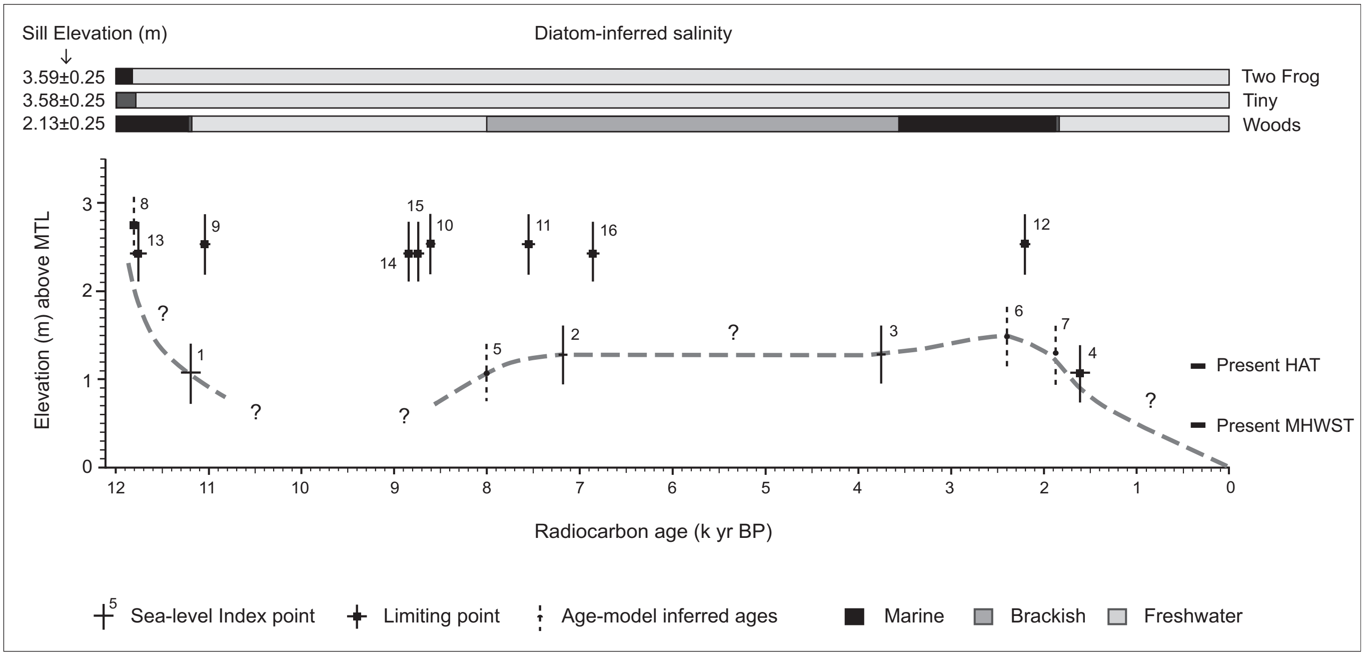

Fig. 7. Late-glacial and Holocene relative sea-level (MTL) curve (^{14}C timescale) for the SBIC. Details of the sea-level index points and limiting dates (when associated RSLs were above the basin sills) are given in Table 1. Note: the age-model inferred points have lower chronological precision and are less reliable than the other points. Tidal datums (HAT and MHWST) are based on information supplied by the Canadian Hydrographic Service (2002).

unknown, although palynological evidence confirms that late-glacial forest communities, dominated by pine and alder, were well developed in the region during the earliest phase of marine silt and clay deposition (Galloway et al., 2007, 2009; Stolze et al., 2007).

The abrupt nature of the isolation contacts at all three sites points to rapid land emergence (Long et al., 1999). The data from Tiny Lake suggest that isolation occurred prior to 11,763 ± 87 ^{14}C BP (13,806–13,406 cal BP), although the corrected basal gyttja date from Woods Lake indicates that this basin was isolated somewhat later, shortly before 11,195 ± 108 ^{14}C BP (13,300–12,774 cal BP). This offset may partly reflect the elevation of the sills; the sill at Woods Lake (2.13 ± 0.25 m MTL) is 1.45 m lower than that at Tiny Lake (3.58 ± 0.25 m). The reservoir correction factor applied to the basal contact date at Woods Lake may also contribute to the later age estimate. (Note: a correction factor was not applied to the lower-most gyttja date (at 3.38 m) for the Tiny Lake core since this sample lay 6 cm above the isolation contact). Palynological evidence from Woods Lake also suggests that marine conditions persisted for longer at this site. In particular, the presence of spruce and mountain hemlock pollen, which place a higher ecological demand on growing conditions (e.g., soil, light) than pine and other early boreal forest taxa, suggest that a more advanced stage in the late-glacial vegetational succession was achieved before isolation (cf. Stolze et al., 2007).

Due to the age reversal in the basal gyttja date from Two Frog Lake (Table 1), the timing of isolation here is less certain. An age-model based estimate for the isolation contact at c. 4.91 m (Fig. 3) suggests that the sea withdrew from the basin at around the same time as at Tiny Lake, at ~11,800 ^{14}C BP.

3.4.2. *Early Holocene*

RSL position during the early Holocene is poorly constrained, although limiting dates from freshwater sediments at Tiny and Two Frog lakes confirm that the sea remained below c. 3.6 m MTL between 8840 and 8620 ^{14}C BP (Fig. 7). The continuous freshwater conditions observed at Woods Lake imply that RSL was <2.13 m MTL throughout the early Holocene.

3.4.3. *Middle Holocene*

The Woods Lake diatom record provides important evidence for a marine transgression during the early middle Holocene that breached the 2.13 m sill of the former basin, but failed to penetrate the ~3.6 m sills of the other two (Fig. 7). A RSL rise to 1.28 ± 0.34 m MTL at 7176 ± 44 ^{14}C BP (8154–7878 cal BP) is indicated (Fig. 7). An age-model derived estimate based on the ingression contact (at 1.65 m) at Woods Lake loosely constrains the onset of this transgression to ~8000 ^{14}C BP (8900 cal BP) (Fig. 3).

The well-developed sand layer in the middle Holocene limnic sediments at Woods Lake is noteworthy. The lateral extent of this unit (Fig. 2) and the concomitant rise in marine and brackish water diatoms, suggest that the sand was introduced via wave or tidal processes rather than by streams, although the size (up to ~1 cm diameter) and orientation of the occasional clasts observed in the core may also suggest derivation from slope-inputs (Appendix A). Aeolian deposition cannot also be ruled out for the finer clastic sediments. Indeed, the presence of intermittent sand layers in the freshwater sediments from the same interval in the other two lakes points to a non-marine origin. It is possible that higher-than-present RSLs during the mid-Holocene resulted in an opening of the vegetation on the coastal side of the lakes, exposing the basins to aeolian inputs (Galloway et al., 2009). The high frequencies of *Aulocoseira* diatoms in the early and middle Holocene parts of the Tiny Lake core (Fig. 6) support this, since the heavily silicified cells of this planktonic taxon require turbulent water conditions to maintain suspension in the photic zone.

Inspection of the age-models for the three lake cores (Fig. 3) suggests that sand input into the basins during the mid-Holocene may have led to local changes in sediment accumulation rates, although it should be noted in the case of Woods Lake that there is no evidence from the stratigraphy for erosion associated with the saltwater ingression (Doherty, 2005). Stolze et al. (2007) speculate that subtle changes in the pollen spectra (notably the Cupressaceae pollen record) might reflect short-lived erosion of the sediment fill around the onset of the marine incursion, although the pollen spectra from Tiny Lake (Galloway et al., 2009) and Two Frog Lake (Galloway et al., 2007) do not deviate significantly, suggesting that the noted differences may reflect local taphonomic controls. If this inference is correct, and uninterrupted sedimentation did prevail at Woods Lake, then brackish to fresh-brackish conditions associated with slightly higher-than-present sea-levels are suggested throughout the entire middle Holocene (Fig. 7).

3.4.4. Late Holocene

A further small rise in RSL at ~2400 ^{14}C BP (2450 cal BP) is suggested by the significant expansion of *P. sulcata* diatom communities observed in the Woods Lake core above 1.00 m (Figs. 4 and 7). The sill at Woods Lake continued to be regularly inundated by tides until ~1850 ^{14}C BP, when, based on the age model, RSL fell to 1.28 ± 0.34 m MTL and saltwater only entered at the basin during the highest tides. Isolation was completed by 1604 ± 36 ^{14}C BP (1565–1404 cal BP).

3.5. Geophysical model predictions of RSL

Fig. 8a shows RSL predictions for Woods Lake for the five viscosity models described in Section 2.5. We also show a prediction

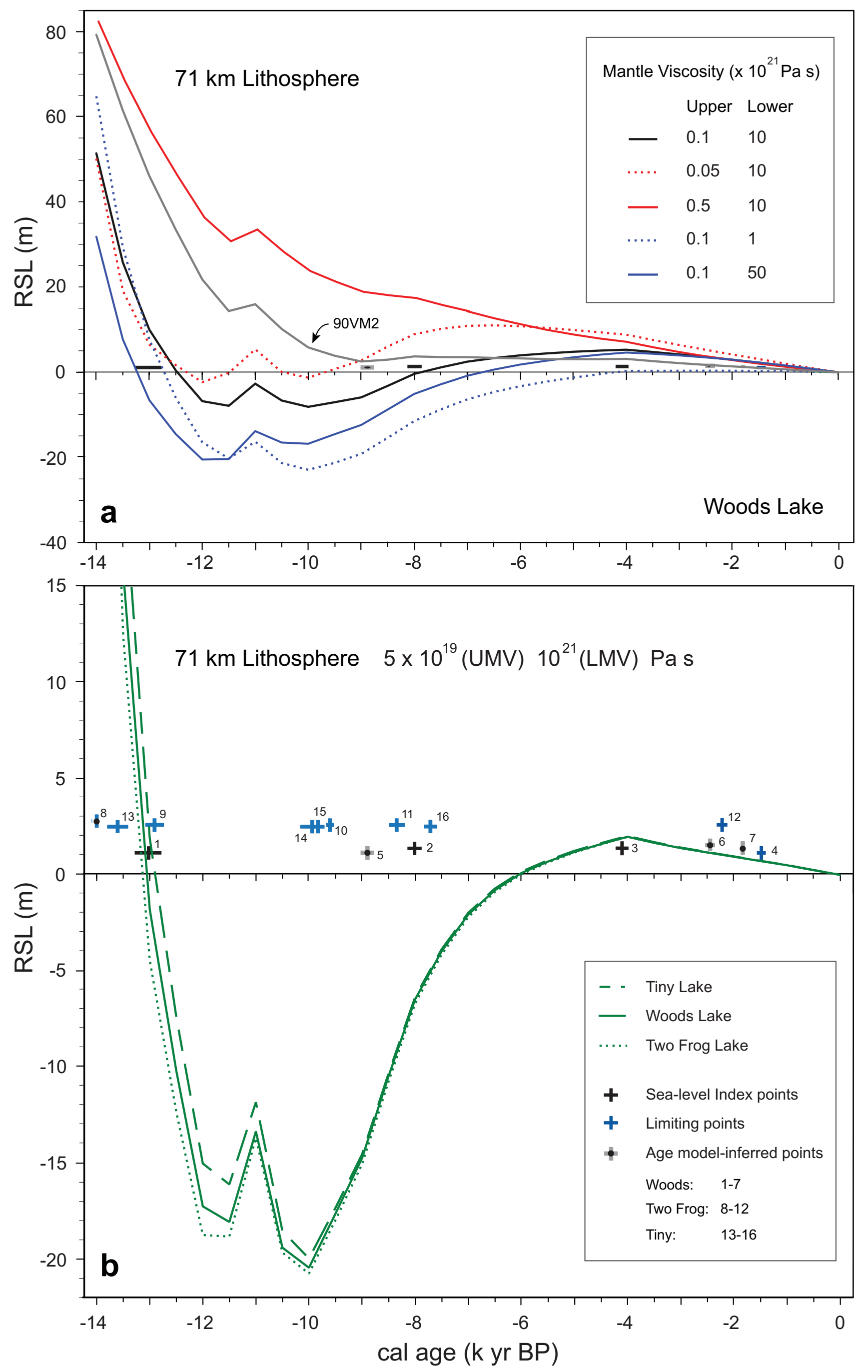

Fig. 8. Model-predicted RSL curves (plotted in cal BP) based on the ICE-5G model, with varying Earth model parameters. Inset (a) shows model predictions (6) and observational RSL data for the Woods Lake basin; (b) shows model predictions at all three lakes for the Earth model with a lithospheric thickness of 71 km, and UMV and LMV values of 5×10^{19} Pa s and 10^{21} Pa s, respectively. Of the small subset of Earth models considered, this one captured most of the data at Woods Lake. Note that observations are also shown for all three lakes. Details of the RSL data points and limiting dates are given in Table 1. Note: the age-model inferred points have low chronological precision.

using an Earth model that ICE-5G is calibrated to (the VM2 viscosity model; in this case with a 90 km thick lithosphere) (Peltier et al., 2002; Peltier, 2004). We focus on this site first because the greatest number of sea-level reconstructions was produced here compared to the two other lakes (Table 1).

The suite of model predictions presented indicates the sensitivity of the simulated RSL curves to changes in UMV and LMV. Inspection of these curves shows that the changes considered for UMV (compared black and red lines) have the greatest impact on the model output at this site. (Very similar sensitivity is evident at the other two sites.) The higher values for UMV considered (5×10^{20} Pa s and those in the many layer VM2 model, which average to around this value in this region) result in a RSL curve that does not drop below present sea level. The variations in LMV considered have a smaller impact on the model RSL curves (compare black and blue lines) as the form of the curve is similar for the (wide) range of values considered.

The first order result from this preliminary sensitivity analysis is that the data are not compatible with UMV values greater than $\sim 10^{20}$ Pa s. This is consistent with earlier studies that considered data to the south of our study area (James et al., 2000, 2009b). Such low values of viscosity are required to produce the rapid rates of sea-level fall indicated by the RSL observations. For the adopted ice model, the sea-level index point at c. 13,000 cal BP (#1; Table 1) can only be captured with UMV values that are 10^{20} Pa s or less. Low UMV values (10^{18}–10^{19} Pa s) are also required to re-produce the large crustal uplift rates observed in southeast Alaska (Larsen et al., 2005), which is in a similar tectonic setting to our study area. The existence of such low values is compatible with geodynamic models of the subduction process (e.g. James et al., 2009b).

Apart from the data requirement for rapidly falling sea levels during the late-glacial, the other key constraints at Woods Lake are a return to greater than present values around 8000 cal BP with a mid- to late Holocene highstand not exceeding ~ 2 m. None of the viscosity models considered is compatible with these three criteria. A more extensive search through Earth model viscosity space is required to determine the combination of UMV and LMV values that are most compatible with the data. This is beyond the scope of this study in which the primary aim is to present the data and a preliminary data-model comparison. However, based on the results shown in Fig. 8a, we did consider a small number of additional models that fell within the range considered and might provide a reasonable fit. Of those, we chose the model with an UMV of 5×10^{19} Pa s and a LMV of 10^{21} Pa s, since it fits quite well the observations during the late-glacial and the late Holocene; the fit in the early Holocene is poor, however (see Fig. 8b).

In Fig. 8b, we show predictions for this Earth model with the data from all three lakes. Apart from the major discrepancy during the early to middle Holocene, it is worth noting that the RSL fall during the late-glacial does not capture the data points from Two Frog Lake (#8) and Tiny Lake (#13). These limiting dates should lie above the predicted RSL curve, suggesting that a less steep RSL fall would improve the fit. The only way to achieve this and capture the Woods Lake data at 13,000 cal BP is to have earlier ice retreat (and a greater value for UMV).

A final feature of the predictions in Fig. 8b that is worthy of note is the offset in RSLs (of c. 1–10 m) between the three basins prior to 10,000 cal BP. Since the lakes lie in a triangle only c. 20 km apart (Fig. 2), this indicates that there is a significant gradient in the RSL response associated with ice-ocean loading. Predicted RSLs for Tiny Lake, which lies furthest inland and closer to the likely centre of ice loading, deviate most significantly. We have insufficient evidence to fully appraise whether this offset is borne out by the RSL data, although our preliminary observations suggest that, contrary to the predictions, the Tiny Lake basin was isolated before Woods Lake.

It is clear that, in order to seek optimal fits to the data, a more comprehensive modelling analysis is required that considers both a greater number of Earth viscosity models as well as revisions to the ICE-5G loading history in this region.

4. Regional comparisons

4.1. Late-glacial

The RSL data presented provide new insights into the character of post-glacial sea-level change in an area of the British Columbia coast that has remained largely uninvestigated and a preliminary basis for appraising the wider regional response of the central mainland coast to deglaciation. The data also provide new constraints on the timing of local ice retreat. The isolation of the Tiny Lake basin by $\sim 11,800$ ^{14}C BP, for example, and the evidence for forested conditions during the period of marine sedimentation that preceded this (Galloway et al., 2009), confirm that the outer part of the inlet complex was ice free well before 12,000 ^{14}C BP. The timing of deglaciation in this area of the mainland coast is not well known, although this age estimate is in agreement with evidence from dated marine shells from Port McNeill on northeastern Vancouver Island, ~ 55 km to the south of the Nakwakto Rapids, which suggest that deglaciation commenced locally before 12,930–12,250 ^{14}C BP (Howes, 1983). To the northeast on the mainland at Bella Coola (Fig. 1), deglaciation is estimated to have occurred later at 10,790–10,420 ^{14}C BP (Andrews and Retherford, 1978). This is consistent with an eastward pattern of glacial retreat towards the interior (cf. Clague, 1981; Clague and James, 2002).

The broad sea-level trends observed in the SBIC in late-glacial times show close correspondence with those recorded in other areas of the mainland coast and Vancouver Island (Fig. 1D–H), although in contrast to some of these areas, the late-glacial marine highstand and the early phase of rapid emergence that followed deglaciation remain unconstrained. It is worth noting though that sea levels appear to have been lower in the SIBC between $\sim 12,500$ and 11,200 ^{14}C BP than at Bella Bella to the immediate north, where well-preserved barnacles from ~ 15 m, reported in the 1970s, yielded a date of 12,210 $\pm$ 330 ^{14}C BP (Lowdon and Blake, 1973; Andrews and Retherford, 1978). If the modern oceanic reservoir correction factor (950 $\pm$ 100 y) that has been applied to late-glacial marine shell dates elsewhere in the region (e.g., Hutchinson et al., 2004b) is applied, a corrected age of 11,260 $\pm$ 330 ^{14}C BP is indicated. This may reflect a differential response to ice loading between the two regions.

A more pronounced diachroneity in the RSL response to deglaciation is observed if the records from the SBIC are compared with those from the north mainland coast. At Kitimat, for example, which lies ~ 100 km inland from the open coast (Fig. 1), ice remained until 9700 ^{14}C BP (Hetherington et al., 2004). RSLs remained high throughout the late-glacial and at $\sim 10,000$ BP stood well above 100 m (Clague et al., 1982; Hetherington et al., 2004; Fig. 1D). On the outer coast, at Prince Rupert (Fig. 1), a more prolonged phase of late-glacial marine inundation is also suggested than in the SIBC by marine shells which lie at 13 m elevation and date from 12,700 $\pm$ 120 BP (Clague et al., 1982) (corrected age = 11,750 $\pm$ 130 BP). This again may reflect later deglaciation in this more northerly region.

A much closer correspondence is apparent when the SIBC data are compared with records from the area to the southeast, particularly eastern Vancouver Island (Hutchinson et al., 2004a). RSLs fell here from 150 m at $\sim 12,500$ ^{14}C BP to present levels shortly before 11,500 ^{14}C BP (Mathews et al., 1970; Clague et al., 1982; Fig. 1F curve a). Isolation basin sequences from the adjacent central Strait of Georgia (Fig. 1) indicate that present sea levels were achieved by

~11,000 ^{14}C BP (Hutchinson et al., 2004a). This is very similar to the observations from the SBIC, where RSLs stood at ~1 m above MTL at ~11,200 BP (Fig. 7). Interestingly, a RSL record from the northern Strait of Georgia, 200 km to the southeast of the Nakwakto Rapids, is the only record from this area to show any deviation. Sea levels here stood at 45 m at 11,000 ^{14}C BP, following an initial period of rapid emergence, and at 10–15 m at 10,000 ^{14}C BP (James et al., 2005; Fig. 1F curve b). A further fall to present levels at ~9700 ^{14}C BP is tentatively inferred (James et al., 2005: Fig. 6). The reason for this offset in the timing of the RSL changes around the Strait of Georgia is unclear, although it may reflect small differences in crustal tilting associated with rapid deglaciation of the Strait area (James et al., 2002).

A more striking contrast is observed between the SBIC RSL data and records from Queen Charlotte Sound and the Hecate Strait to the immediate northwest (Hetherington et al., 2004; Fig. 1A–C). These areas lie within an inferred forebulge zone, which is thought to have persisted from ~13,200 ^{14}C BP until after 9700 ^{14}C BP (Hetherington and Barrie, 2004). The interplay of low eustatic sea levels and crustal bulging meant that shorelines were significantly lower than present here in late-glacial times, although the RSL response was complex and varied spatially and temporally as the forebulge migrated to the east during deglaciation (Clague et al., 1982; Hetherington and Barrie, 2004; Hetherington et al., 2004). The significant difference in shoreline elevations (>130 m) between Queen Charlotte Sound at 12,000 ^{14}C BP (as indicated in the sea-level curve of Hetherington and Barrie, 2004; Fig. 1C), and the data from Tiny Lake (Figs. 1 and 7) is remarkable and points to steep crustal tilting between the two regions.

4.2. Late-glacial–early Holocene transition

The SBIC data provide no insights into shoreline position between 11,800 and ~8000 ^{14}C BP, as sea levels lay below the sills of the three basins during this time. The inferred lowstand is nevertheless similar in timing to that observed in many areas of the southeastern British Columbia coast and probably reflects the combined effect of low eustatic sea levels and diminished rates of isostatic recovery (cf. Clague et al., 1982; Clague and James, 2002; Hutchinson et al., 2004a; James et al., 2009a). The timing of the lowstand is similar, for example, with that reported in eastern Vancouver Island and the central Strait of Georgia (Clague et al., 1982; Hutchinson et al., 2004a). Dated isolation basin sequences from the latter area indicate that sea levels returned to present levels at ~8800–7900 cal BP [~8300–8000 ^{14}C BP] (Hutchinson et al., 2004a). This overlaps with the age-model derived estimate for the re-establishment of brackish water conditions at Woods Lake (8000 ^{14}C BP) (Fig. 7).

Sea-level evidence in the Bella Bella–Bella Coola region to the north is sparse for this interval. In contrast to the SIBC though, the sea is thought to have fallen below present levels between 9000 and 7000 BP and remained a few metres below present until the late Holocene (Clague et al., 1982; Fig. 1E).

The RSL changes during this period are more fully constrained in Queen Charlotte Sound (Fig. 1C). Sea levels here were at least 95 m lower than present between 10,650 and 10,000 ^{14}C BP (Luternauer et al. 1989). Similar elevations have been inferred from dated organic remains in cores from the southern Cook Bank area, only 70–80 km west of the Nakwakto Rapids (Hetherington et al., 2004; Figs. 1C and 2). Hetherington et al. (2004) argue that a large coastal plain developed in this region between ~14,000 and 9700 ^{14}C BP. This is thought to have extended to northern Vancouver Island and occupied parts of the shelf to the immediate west and north of the SBIC (Hetherington et al., 2004: Fig. 8). They speculate that this plain may at some point have extended across to the mainland, forming a landbridge that facilitated faunal migrations into Vancouver Island. The presence of ~12,000 year old skeletal remains of mountain goat (*Oreamnos americanus*) in caves in northern Vancouver Island (Nagorsen and Keddie, 2000) adds weight to this hypothesis. If such a significant RSL fall did occur in this part of the central mainland coast and adjacent shelf, then it is possible that the SBIC may have become completely isolated, as the sill beneath the Nakwakto Rapids lies at only −35 m (Carlson and Hobler, 1976). The early Holocene freshwater diatom assemblages and palynological evidence from the SBIC basins unfortunately shed no further light on this, nor do marine sediment cores collected from the floor of the inlets, as they only extend back to the mid- to late Holocene (Patterson et al., 2007; Vázquez-Riveiros and Patterson, 2009; Galloway et al., 2010; Babalola et al., 2013).

4.3. Middle–late Holocene

The evidence for a prolonged phase of slightly higher sea-levels in the SBIC between ~8000 and ~1850 ^{14}C BP (Fig. 7) is a prominent characteristic of the Holocene RSL record that requires explanation. In considering wider evidence for similar trends, it should be noted that few records of sufficient precision exist to accurately constrain such a small highstand, as the associated heights (up to ~1.50 m above present MTL), lie well within the range of extreme tides or storm deposition. Notwithstanding this, higher-than-present sea levels have not been reported from the north mainland coast or the Bella Bella region during this time (Andrews and Retherford, 1978; Carlson, 1979; Clague et al., 1982; Fig. 1D and E); here dated terrestrial deposits and archaeological remains indicate that the sea remained slightly below present levels. Nor have they been observed in southern areas of Vancouver Island and the Fraser Lowland (Clague et al., 1982; James et al., 2009a; Fig. 1G and H). Higher RSLs have nevertheless been inferred at several locations in northern, eastern and western Vancouver Island from dated raised beach and other intertidal deposits (Hutchinson, 1992; Hutchinson et al., 2004a). The timing and magnitude of these changes are poorly constrained, however. At Comox, in eastern Vancouver Island, for example, the height of the mid-Holocene transgression is constrained by a shell date underlying tidal sand deposits that extend up to 2 m above present datum (Hutchinson et al., 2004a). This has yielded a corrected age of 5690 ± 120 ^{14}C BP (Hutchinson et al., 2004a). Evidence from other sites in this area suggests that RSLs rose above present around 8000 BP and the upper tide limit remained at ~4 m above present until ~3000 BP (Hutchinson, 1992). Isolation basin records from the adjacent central Strait of Georgia also indicate that RSL rose to ~1 m above present levels at ~8300–8000 ^{14}C BP before slowly declining (Hutchinson et al., 2004a). Coastal basin records from the northern Strait of Georgia 80 km to the north, unfortunately do not have good precision for the middle Holocene, although sea level stood at, or slightly below, 1.5 m at 2300–2000 ^{14}C BP before declining to present levels (James et al., 2005; Fig. 1F, curve b).

Closer to the SBIC in northern Vancouver Island, higher mid- to late Holocene sea levels have also been reported on the Brooks Peninsula (Hutchinson, 1992; Fig. 1). The evidence here is based on dated raised beaches and intertidal sediments found in relict sea caves that lie ~3–4 m above present high tide. These records suggest that higher-than-present sea levels persisted from at least the middle Holocene until 700 ^{14}C BP (Hutchinson, 1992). The causes of this spatially extensive yet locally variable phase of high sea levels are unclear, although they probably reflect differential rates of residual crustal recovery operating across the region. The fact that more northerly areas of the mainland coast areas remained submerged during most of this time may reflect a later age of deglaciation, or some peripheral influence of residual forebulge collapse in the adjacent Queen Charlotte Sound Basin (cf. Hetherington and Barrie, 2004).

A final feature in the SIBC record of interest is the salinity increase inferred at Woods Lake between ~2400 and 1900 ^{14}C BP, which we speculate may reflect a small RSL rise. Without a wider degree of coherency in the regional late Holocene RSL records, it is difficult to interpret this. It is nevertheless interesting that it coincides in timing with Neoglacial glacier advance in the Coast Mountains near Bella Bella at ~2500 ^{14}C BP (Desloges and Ryder, 1989). This corresponds to the Tiedemann Advance reported in the southern British Columbia Coast Mountains (Ryder and Thomson, 1986). The pollen spectra from three SBIC basins certainly show evidence for cooler, moister conditions after c. 3900 ^{14}C BP (Galloway et al., 2007, 2009; Stolze et al., 2007) like many other records from coastal British Columbia (Hebda, 1983; Lacourse, 2005). It is possible that the Woods Lake basin may have lain in a sufficiently sensitive recording position (just above mean tide level) to record small-scale RSL rise associated with crustal loading in the hinterland during this period of cooling and ice advance. This inference is speculative though and requires further testing, as it remains to be seen whether Neoglaciation could have generated a RSL response in this region. A more elaborate interplay of long-term residual isostatic adjustments, coupled with eustatic change may provide a better explanation for the observed RSL changes.

4.4. Summary

The close correspondence shown between the SBIC RSL data and the more complete records from areas to the immediate southeast for the same interval, particularly the central Strait of Georgia and eastern Vancouver Island, suggest that the deglacial chronologies of these regions were broadly similar. The observed discrepancies noted with the data from the Bella Bella region and other outer coast areas on the north mainland coast may reflect greater ice-loading or a later timing of deglaciation in the latter regions. This would be in line with the broad pattern of ice retreat, which was towards the north and the interior following initial deglaciation from the west (Clague, 1981; Clague et al., 1997; Clague and James, 2002). The differences may also reflect intra-regional isostatic controls which cannot currently be elucidated from the available RSL datasets (e.g., Andrews and Retherford, 1978; Clague et al., 1982).

5. Conclusions

- This study is the first to reconstruct post-glacial RSL change in the central mainland BC coast (Seymour-Belize Inlet Complex) using dated sediments from isolation basins. Twelve new sea-level index points have been described.
- Three low-lying basins in the outer part of the SBIC were flooded by the sea in late-glacial times and subsequently isolated between ~11,800 and 11,200 ^{14}C BP. Previously published palynological data suggests that forested conditions were well established locally before ~12,000 ^{14}C BP, confirming that the area was ice free before this time. Between ~11,200 and ~8000 ^{14}C BP, RSLs fell below the sill (2.13 ± 0.25 m above present MTL) of the lowest basin, Woods Lake, and freshwater conditions prevailed.
- The Woods Lake record provides important insights into the timing and extent of Holocene RSL change, showing that sea levels stood at around ~1.30–1.50 m above present MTL between ~8000 ^{14}C BP until ~1900 BP. The sea did not, however, penetrate two slightly higher coastal basins (sills heights = ~3.6 ± 0.25 m above present MTL) during this interval. RSLs fell to ≥1.07 m MTL by 1604 ± 36 ^{14}C BP, isolating the Woods Lake basin for a final time.
- The RSL history of the SBIC shows several similarities with the more complete records from eastern Vancouver Island and the central Strait of Georgia to the southeast of the region. A comparable pattern of ice loading and deglaciation may tentatively be inferred. The record shows some differences, however, with RSL curves from the mainland coast near Bella Bella and Prince Rupert, suggesting that the crustal response to deglaciation differed to these more northerly, outer coast areas. This may reflect enhanced ice loading and/or later timing of deglaciation. Further detailed comparisons are hindered by a lack of precision in the available RSL datasets from these regions and the general paucity of data for some intervals, particularly the late-glacial period.
- Model predictions generated using the ICE-5G model partnered with a small number of different Earth viscosity models generally show poor agreement with the observational data, indicating that the ice model and/or Earth models considered can be improved upon. The best data-model fits were achieved with relatively low values of upper mantle viscosity (5×10^{19} Pa s), which is consistent with previous modelling results from the region.
- The value of using isolation basins to constrain RSL change has been clearly demonstrated and the application of this approach in this and adjoining regions should be investigated further (cf. James et al., 2002; Hutchinson et al., 2004a).

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Appendix A. Supplementary data

Supplementary data related to this article can be found at http://dx.doi.org/10.1016/j.quaint.2013.01.026.

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