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On the postglacial isostatic adjustment of the British Isles and the shallow viscoelastic structure of the Earth

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SUMMARY

Observations of postglacial relative sea level (RSL) history at sites beyond the margins of the main accumulations of Würm–Wisconsin–Devensian ice may be invoked to test the accuracy of global models of the glacial isostatic adjustment process. The existing database of RSL time-series for the British Isles forms a continuous network of sites that sample the coasts of Scotland, England and Wales fairly uniformly, and which may be especially important in this regard. Not only are these ^{14}C data for the Holocene epoch of very high quality in terms of our knowledge of the coastal environment and relationship to palaeo-sea level of each age measurement, but they are especially numerous and relate to an island that was itself partly glaciated in the north, essentially ice-free in the south, and located well outboard of the massive ice-sheet that covered Fennoscandia at the Last Glacial Maximum (LGM).

Taken together, the time-series from 55 distinct locations allow us to perform a fairly stringent test of the recently formulated ICE-4G (VM2) global model of the isostatic adjustment process. Even without modification, this model has been shown to fit most of the time-series in the UK data set rather well. In order to satisfy very recently published *a priori* constraints on the maximum thickness of the Scottish ice-sheet, however, we find that the lithospheric thickness, the feature of the radial viscoelastic structure to which the rebound data from Scotland are most sensitive, must be somewhat reduced. Whereas the assumed thickness of the lithosphere in ICE-4G (VM2) was near to 120 km, best fits to the Scottish data require a lithospheric thickness near to 90 km. This reduced value for lithospheric thickness agrees exceedingly well with the latest estimates of this Earth property based on the inversion of oceanic geoid and topography data, as the asymptotic thickness characteristic of the oldest ocean floor. This would appear to be the best possible *a priori* estimate for the thickness of the cratonic lithospheres upon which the largest ice-sheets of the late Pleistocene are known to have rested. We also demonstrate explicitly that inferences of the shallow viscoelastic structure beneath the British Isles are strongly influenced by the details of the history of deglaciation at remote locations, at Antarctica in particular.

Key words: continental lithosphere, glacial isostasy, sea level.

1 INTRODUCTION

The gravitationally self-consistent global viscoelastic theory of the glacial isostatic adjustment (GIA) process, most recently reviewed in Peltier (1998a), now serves as an important and widely employed cornerstone of geodynamics research. This is a consequence of the fact that, through the use of this theory to rigorously invert the wide range of geophysical observations that are now known to be related to the GIA process, one may make useful inferences of important characteristics of the radial viscoelastic structure of the planet and of the surface deglaciation process that occurred subsequent to the LGM. Although most of the work in this area has focused on the inverse problem for mantle viscosity (see the papers by Peltier 1998b and Wieczerkowski *et al.* 1999 for recent examples), much less effort has been expended on attempts to infer the properties of the near-surface viscoelastic structure, the most significant feature of which is the thickness of the 'lithosphere' in which the effective viscosity is so high that the behaviour of this layer is essentially perfectly elastic. Since the isostatic adjustment of the Earth subsequent to the removal of a large spatial-scale ice-sheet from its surface (such as the Laurentide ice-sheet which was centred on Hudson Bay and essentially covered all of Canada at the LGM, or the Fennoscandian ice-sheet which was then in place over Norway, Sweden and Finland) is only weakly dependent upon lithospheric thickness, if we wish to attempt to measure lithospheric thickness through the inversion of GIA data we are obliged to focus upon the response of the planet to the removal of loads of smaller spatial scale.

It is for this reason that observations of the postglacial recovery of the British Isles are expected to be most useful from a geodynamics perspective, especially because the small Scottish ice-sheet rested upon the same Precambrian terrain as did the larger Fennoscandian and Canadian ice-sheets. (Clearly, the response to the removal of a small-scale load such as that which covered Iceland at the LGM, because of its location over a mid-ocean ridge, would be expected to sense the existence of a much thinner lithosphere capping a much less viscous upper mantle than is expected to be characteristic of cratons.) As we will discuss fully in Section 2, the existing data set for Great Britain is for the most part entirely consistent with expectations based upon a previously constructed model of the GIA process denoted ICE-4G (VM2). The radial viscoelastic structure of this model, denoted VM2 for Viscosity Model 2, was originally inferred (Peltier 1996a,b, 1998b; Peltier & Jiang 1996) on the basis of a formal Bayesian inversion of a very small and appropriately parametrized subset of the global database of glacial isostatic adjustment observations that was compiled in Toronto, beginning with the reconnaissance collection of RSL time-series described by Tushingham & Peltier (1991). Rather than being based upon the 'raw' age–height pairs, this reconnaissance collection included many time-series constructed by sampling the 'envelope' of the distribution of age–height pairs, as in the early analysis by Walcott (1970; 1972) of the ^{14}C database of RSL histories that was then available from sites on the Laurentian platform of Canada. Later versions of the database have been compiled by recording the raw age–height pairs themselves, and it is this type of data that will be employed herein.

The most significant argument favouring the ICE-4G (VM2) model over its ICE-3G (VM1) precursor (Tushingham & Peltier 1991) was that it essentially eliminated the very large misfits predicted using the earlier model of RSL histories to the relative sea-level observations from all sites along the eastern seaboard of the continental United States (see, for example, Peltier 1996a and Fig. 23 in Peltier 1998a). Tushingham & Peltier (1992) had identified this region as the one in which the misfits to the predictions of their simple ICE-3G (VM1) model were most severe. Because none of the US east-coast data had been employed in the construction of ICE-4G (VM2), the fact that these misfits were very significantly reduced by this model is construed to have provided a clear demonstration of the improvement to the viscosity structure embodied within it (the differences between the ICE-4G and ICE-3G Late Pleistocene deglaciation histories were slight).

In both the VM1 and VM2 radial viscoelastic structures, the surface lithosphere was assumed to have a thickness of 120.7 km. The basis of this assumption, which will be further tested herein through analysis of postglacial RSL observations from the British Isles, lay in the early work by Parsons & Sclater (1977) involving investigations of the variation of ocean-floor bathymetry and gravity as a function of age. On the basis of an admittance function analysis, these authors had inferred the asymptotic thermal thickness of the oceanic 'plate' to be 125 ± 10 km, with this asymptotic thickness being realized beneath ocean floor of maximum age. In constructing the VM1 and VM2 models this estimate was employed directly, the assumption being that the thickness of the continental lithosphere on which the LGM ice-sheets were located should be approximately equal to this thickness. In the very recent past (DeLaughter *et al.* 1999), the Parsons and Sclater analysis has been considerably extended and refined, leading to the conclusion that the asymptotic thermal thickness of the oceanic lithosphere is in fact considerably smaller than this earlier estimate, and best approximated by a thickness of 95 ± 10 km. In the present paper one goal will be to test the plausibility of this revised estimate on the basis of the analysis of the GIA observations available from the British Isles.

In past work on the inference of lithospheric thickness using data related to the GIA process, results have always proved to be somewhat equivocal. Peltier (1984; 1986), for example, attempted to infer the magnitude of this parameter on the basis of the tilts of postglacial shorelines of Holocene age that are inscribed into the landscape in the Great Lakes region of the North American continent. He concluded on this basis, and on the basis of early analyses of US east-coast RSL data, that lithospheric thickness could be as great as 200 km or more. In Peltier (1996a,b), however, models of this kind were abandoned when it became clear that appropriate adjustments to the immediately sublithospheric viscosity profile, based upon the requirement that the structure adequately predict the relaxation time spectrum (at longest wavelengths) for the postglacial rebound of Fennoscandia that had earlier been inferred by McConnell (1968), enabled the ICE-4G (VM2) model thereby obtained to reconcile the entire set of US east-coast RSL observations much better. As very clearly demonstrated through the recent reanalysis of the Fennoscandian relaxation spectrum by Wieczerkowski *et al.* (1999), the relevant short-wavelength components of this spectrum provide no significant constraint on lithospheric thickness as their relaxation times are not accurately constrained by the observations. In order to measure lithospheric thickness using GIA observations, information is required that is related to the isostatic rebound associated with an ice load of much smaller spatial scale than either the Laurentide or Fennoscandian ice-sheets. Fig. 1 shows a perspective view of the Earth at the LGM which illustrates the relative spatial scale of the ice-sheet on the British Isles compared to those which were then in place over Canada and northwestern Europe. The former was clearly extremely small in comparison with its more voluminous LGM companions.

The glacial isostatic adjustment of the British Isles has been extensively studied over the course of the past decade, primarily by Lambeck and co-workers (Lambeck 1993a,b, 1995; Lambeck *et al.* 1996), the data being interpreted as preferring a radial variation of viscosity that is adequately approximated by a two-layer structure such that the average viscosity through the upper mantle and transition zone is approximately 0.4×10^{21} Pa ^{14}Cs, whereas the average viscosity in the lower mantle is 10^{22} Pa ^{14}Cs. The adoption of the radiocarbon timescale in these studies, for the purpose of comparing geological observations with theoretical predictions, is a significant difference from the approach adopted here, the implications of which we will discuss further in Section 2. Nevertheless, the viscosity structure that has been inferred on the basis of this previous work is broadly similar to VM2 through the upper mantle and transition zone, although significantly more viscous in the lower mantle in which the VM2 model has an average viscosity of approximately 2×10^{21} Pa s. As has been recently reconfirmed

(Peltier 1998a, see colour Plate 7; Dyke & Peltier 2000), a model of the kind preferred by Lambeck and co-workers, which has a lower mantle viscosity somewhat in excess of 10^{22} Pa, does not fit the observed relaxation times that are characteristic of the postglacial isostatic adjustment of the Laurentide platform of Canada that occurred following the deglaciation of the North American continent. (In particular, a model of this type does not fit the relaxation time characteristic of the rebound process in southeast Hudson Bay, which was a location of the greatest thickness of Laurentide ice; see Fig. 6 in what follows for a plot of relaxation times for RSL data from this geographical region.) Relaxation-time data from sites that were located under the centre of the Laurentide ice-sheet constrain the average viscosity in the upper part of the lower mantle to a value that appears to be bounded above by 2×10^{21} Pa s, the value assigned to the entire lower mantle of the Earth in the simple VM1 model (e.g. see Peltier 1996a). Given the small scale of the ice mass that covered the UK, it should be expected that the relaxation process associated with local GIA would not be sensitive at all to lower mantle structure. The results of the analyses to be described herein will directly demonstrate that the UK data are entirely consistent with the lower mantle viscosity structure that is required to reconcile the Canadian observations. Even the rebound associated with the larger-scale Fennoscandian ice-sheet is only weakly sensitive to the viscosity in the lower mantle of the Earth (Mitrovica & Peltier 1991; Peltier 1996a), meaning that any reasonable value for this Earth property in the upper part of the lower mantle below 660 km depth would be expected to be found acceptable by observations from this region.

Concerning the inference of lithospheric thickness on the basis of rebound data from the UK (or from any other previously glaciated region of similar spatial scale), the primary difficulty encountered in the past has always been to overcome the non-uniqueness that arises because the response to an ice-load of any scale depends not only upon the radial viscoelastic structure but also upon the thickness of the ice that was removed from the surface during deglaciation and which was responsible for inducing the observed rebound of the crust. Recently obtained *a priori* constraints on the maximum thickness of ice that covered the Scottish highlands (Ballantyne 1997, Benn 1997; Ballantyne *et al.* 1998) help to overcome this source of non-uniqueness. In a very recent analysis, Shennan *et al.* (2000a) invoked these constraints and also made use of new RSL observations from northwest Scotland. They tested various combinations of earth- and ice-model parameters and illustrated the variations in the characteristics of the solution that may arise even for sites that are as close to one another as 50–100 km. They obtained the best overall agreement with an earth model that has the same parameters as that described by Lambeck *et al.* (1998), with a lithospheric thickness of 65 km, approximately half the value in VM2 and only two-thirds of the value suggested by the asymptotic thickness of the oceanic lithosphere under the oldest ocean floor according to the recent analysis of DeLaughter *et al.* (1999).

Given our expectation that lithospheric thickness beneath the oldest ocean floor should provide a reasonable estimate of the thickness of cratonic lithosphere, this thin-lithosphere result is clearly problematic. A primary purpose of our present analysis is to employ the recently greatly expanded RSL database from Great Britain to assess the robustness of a 'thin-lithosphere' earth model. If this previous conclusion, that the lithosphere must be thin, were to be reinforced by such further analysis, there may be important consequences for the inferred thicknesses of other LGM ice-sheets of similar scale to that which covered the British Isles, thicknesses that have been determined on the basis of the viscoelastic 'weighing' of these structures using the procedure described in Peltier (1994), and perhaps also for the inferred thickness of the somewhat larger-scale Fennoscandian ice-sheet itself. Since these inferred ice-sheet thicknesses, once isostatically adjusted, map directly into the topography of the planet relative to sea level at the LGM, and since the LGM topography embodied in the ICE-4G (VM2) model, discussed in detail in Peltier (1994, 1996a), has become a key element in the reconstructions of LGM climate currently being performed using coupled atmosphere–ocean general circulation models (e.g. see Vettoretti *et al.* 2000), the 'downstream' geodynamic and palaeoclimatological implications of such a marked reduction of lithospheric thickness could be consequential indeed.

The plan of the paper is therefore as follows. In Section 2 we will fully review the state of the newly constructed UK database of relative sea-level histories, as this has been significantly expanded in the context of the recently completed Land Ocean Interaction Study (LOIS) project (see, for example, Shennan *et al.* 2000). Our focus will be on the discussion of those particular attributes of the sea-level index points that are most significant from the perspective of constraining large-scale models of the isostatic adjustment process. In Section 3 of the paper we will provide an overview of the global theoretical model of the glacial isostatic adjustment process, whose predictions we will employ to compare with the relative sea-level observations described in Section 2. The model requires two primary inputs in order to allow the prediction of local RSL histories, namely a global model of the history of deglaciation from the LGM to present, and a model of the radial variation of viscosity from the Earth's surface to the core–mantle boundary. These inputs will also be discussed in Section 3, as will certain overall characteristics of the expected response of the Great Britain region to post-LGM global deglaciation, including the existence of the land-bridge that joined Britain to the European mainland at the LGM and the history of its inundation. Detailed discussion of the comparison of observed and predicted RSL histories at all sites from the British Isles is provided in Section 4, in which our focus will primarily be upon the issue of the quality of the constraint that these data provide on the shallow viscoelastic structure of the Earth. Considerable attention, however, will also be given to the way in which plausible variations of the deglaciation history at remote locations might be expected to impact the signals observed in the British Isles data set and thereby influence our conclusions concerning the viscoelastic structure. As we will see, these considerations turn out to be important. A summary of the conclusions to which we have been led on the basis of these investigations is offered in Section 5, together with remarks regarding outstanding issues.

2 POSTGLACIAL SEA-LEVEL HISTORIES FROM THE BRITISH ISLES

The post-LOIS database of relative sea-level histories at British Isles locations consists of ^{14}C-dated samples that may be appropriately assigned to the 55 locations indicated on the map in Fig. 2. On this map, each of the indicated sample sites has been

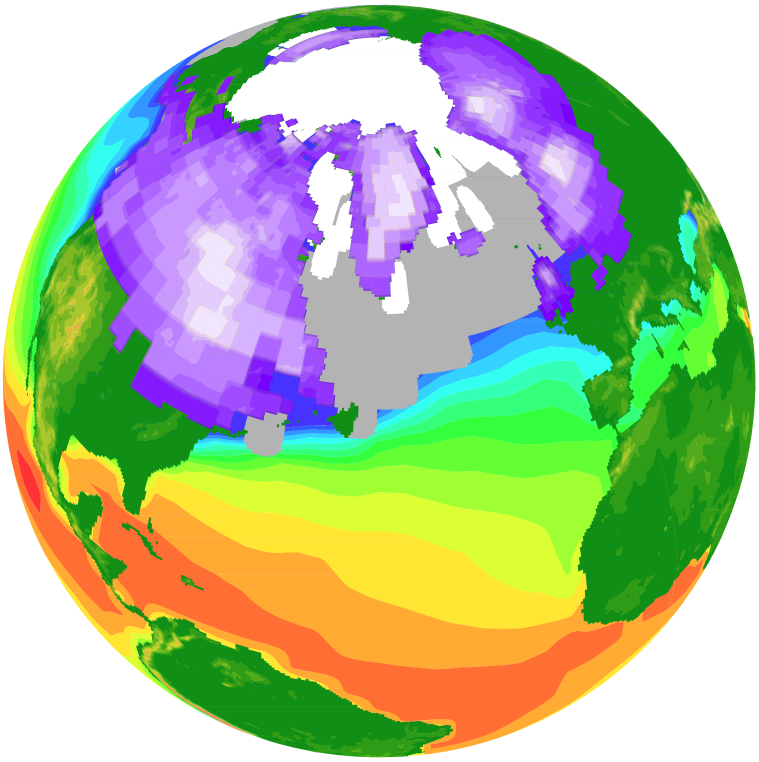


Figure 1. A perspective view of the Earth at the LGM, approximately 21 000 yr BP, showing the parts of the continents that were glaciated at that time according to the ICE-4G model of Peltier (1994, 1996a). Also shown are the summertime (white) and wintertime (grey) regions of the oceans that are expected to have been covered by sea ice. Outside the region of sea-ice cover, the colours over the ocean basins depict annually averaged sea surface temperatures according to the recent global climate model-based reconstruction of Vettoretti *et al.* (2000). Note the relative spatial dimensions of the Laurentide (N. American), Fennoscandian (NW European) and British Isles ice complexes which were in place at the LGM. The Greenland complex did not, of course, disappear during the deglaciation event that commenced subsequent to the LGM.

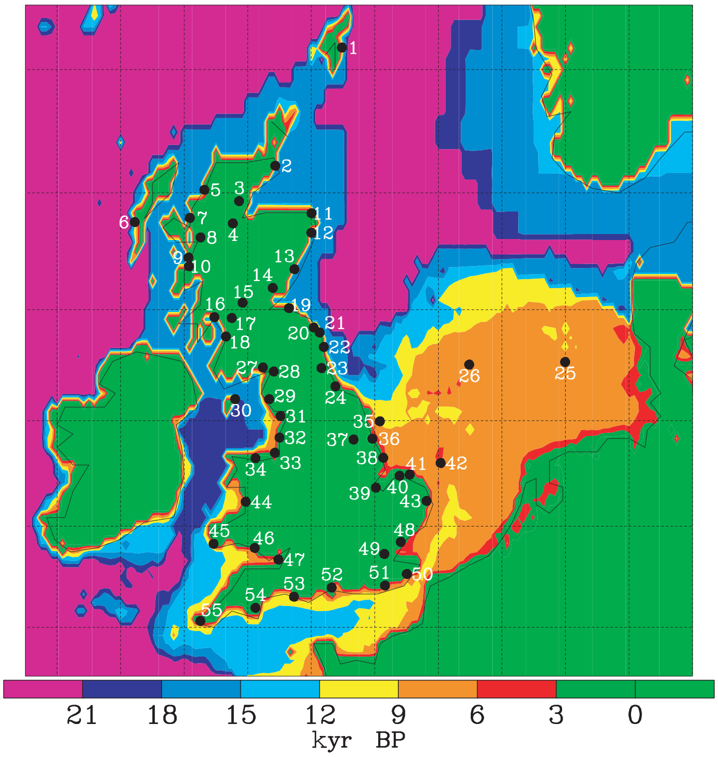


Figure 2. Location map for the 55 ^{14}C-dated time-series that constitute the RSL data base for the United Kingdom. The precise longitude and latitude co-ordinates for each of these sites are listed in Table 1, together with the names that will be employed for reference in this paper. The colour-coded regions that cover the surface of the ocean in the area surrounding the British Isles represent the migration subsequent to the LGM of the coastline separating land from sea in this geographical region. According to the preferred model with lithospheric thickness of $L=90$ km that is inferred on the basis of the analyses described in this paper, Great Britain and France were joined by a vast land-bridge at the LGM, as were Great Britain and Ireland. The time-dependent inundation of these land-bridges that occurred as global deglaciation proceeded will be clear on the basis of inspection of this Figure.

assigned a number. Also shown on this location map, through the colour overlay on the ocean immediately surrounding the British Isles, is the variation of the land–sea distribution that occurred in this region of northwestern Europe from the LGM to present according to the model of glaciation history and radial viscoelastic structure that the analysis to follow will lead us to prefer. The individual time horizons on the map, denoted by the abrupt changes in colour to which distinct ages are assigned on the colour bar, represent the time-dependent position of the coastline, which at the LGM was well outboard of its present location such that an extensive land-bridge then existed connecting Britain to the European mainland and to Ireland. This land-bridge was gradually inundated by the sea as sea level rose during the global deglaciation process that began subsequent to the LGM.

Table 1 presents the complete list of numbered sites from which RSL data are available. In this table we have also provided the full place names for this set of locations, their longitude and latitude co-ordinates, the reference(s) in which the original data are described or summarized, and the number of radio-carbon-dated samples that make up the individual time-series. (In the 'No. data' column the integers N + M refer, respectively, to the number of Mean Sea Level data and 'limiting dates' contained in the database for the site in question. The meaning of the term 'limiting date' is discussed in what follows.) For the purpose of the analyses of these data, the ^{14}C age of every sample in every time-series has been transformed to calendar age using the CALIB 3.0 software of Stuiver & Reimer (1993). Although this has been argued to be unnecessary by Lambeck (1998), we take the view that classical physical theory makes predictions on a uniform 'Newtonian' timescale, not on ^{14}C time. The transformation from ^{14}C years to sidereal years is not linear, as demonstrated explicitly in Fig. 3. We will therefore compare observations with theory on the sidereal timescale and will discuss in later sections the important role that the choice of timescale may play in explaining the differences between the Earth properties we infer by fitting the data and those inferred in the works cited above.

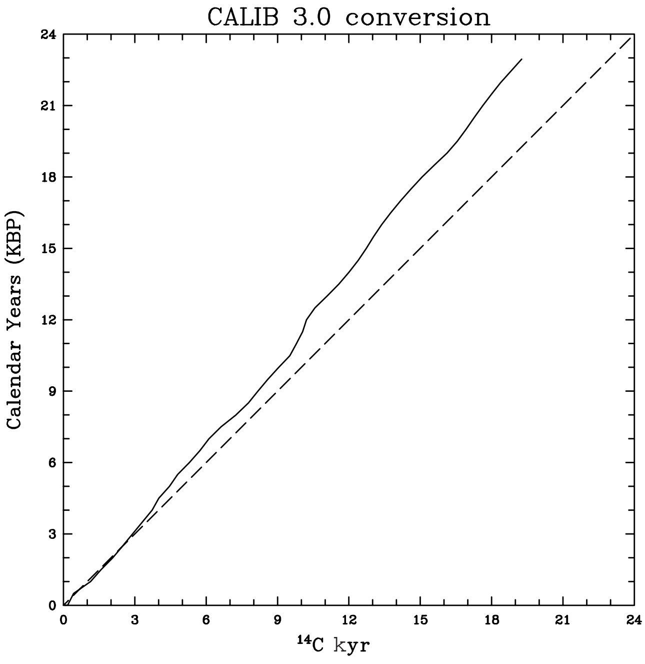


Figure 3. The calibration of the ^{14}C timescale of Stuiver & Reimer (1993) (CALIB 3.0) that is employed for the purpose of the analyses discussed in this paper.

Before discussing the theory that will be employed to reconcile the observations, it will prove useful to comment in some detail on the nature of the relative sea-level data themselves, since each of the time-series contained in the RSL database comprises measurements (^{14}C age and height above present-day sea level) on samples whose indicative meaning in the landscape as regards palaeo-sea level differs from sample type to sample type.

These observations of past sea levels come from a wide range of sources, primarily research articles, the data from the majority of which have not been included in any previous analysis of differential isostatic adjustment on the spatial scale of the entire British Isles or on the timescale of 10–20 ka. Most of these previous studies were focused on geographically restricted areas (for example a single estuary or bay of the order of 10s to 100s of km^2 in area) or on issues related to environmental change that may or may not have included an analysis of relative sea-level history. Application of a consistent methodology to the analysis of such data, however, enables us to extract 'sea-level index points' (henceforth SLIs) from different palaeo-coastal environments and diverse geographical regions. The methodology that is employed for identifying SLIs can be applied universally, to the types of SLIs found on mid-latitude temperate coasts, arctic coasts, or to corals or mangroves from the tropics. This now universally applied approach to the reconstruction of sea-level history was first formalized during International Geological Correlation Programme (IGCP) Project 61, which operated in the period 1979–1983, and has been a component of all subsequent IGCP Projects, especially Projects 200 (Shennan 1983a), 274 and 367. Because of the increasing attention that is being paid to the understanding of such data in the area of solid earth geophysics, and because the interpretation of such data is central to the present paper, our intention in this section is to discuss the methodology thoroughly.

Reliable observations of past changes in sea level relative to the present level derive primarily from sediments, both organic and minerogenic, but also from morphological features whose origin was controlled by palaeo-sea level. To be useful, the sediments must not have been eroded or transported since the time of accumulation or formation. Depending on the nature of the coastal environment, most will record both the vertical and horizontal components of sea-level change; for example, a sea-level rise causes the transgression (migration of the coastal zone in a landward direction) of a low-gradient palaeo-land surface such as a continental shelf. As a result of the processes of weathering, erosion and transport, preservation of sediments and morphological features is limited, with such features having a greater chance of survival in locations sheltered from high-energy geomorphic processes such as storm waves, rivers and tidal channels. Where such sediments and morphological features survive they can be used as SLIs once four attributes have been defined, namely location, age, altitude, and tendency. These form the basis of the database for Britain, and for other areas, that have been developed from the principles established during the series of previously mentioned IGCP Projects (see Preuss 1979; Shennan 1983, 1987a, 1989a,b; Van de Plassche 1986).

The database for Britain comprises over 80 fields of information for each sample, with only a subset of the fields relevant

Table 1. Sources for RSL observations from the British Isles.

Site no.	Lat.	Long.	No. data	Site name	Reference(s)
01	60.343	−1.026	4	Shetlands	Hoppe (1965)
02	58.451	−3.124	11+3	Wick	Dawson & Smith (1997)
03	52.864	−4.263	7	Dornoch Firth	Smith *et al.* (1992)
04	57.490	−4.460	9	Moray Firth	Haggart (1987); Firth & Haggart (1989)
05	58.053	−5.356	7+1	NW Scotland (Coigach)	Shennan *et al.* (2000a)
06	57.512	−7.547	2	Hebrides	Ritchie (1985)
07	57.582	−5.814	7	NW Scotland (Applecross)	Shennan *et al.* (2000a)
08	57.251	−5.476	11	NW Scotland (Kintail)	Shennan *et al.* (2000a)
09	56.907	−5.854	36	NW Scotland (Arisaig)	Shennan *et al.* (1993, 1994, 1995b, 1999, 2000a)
10	56.757	−5.842	14+1	NW Scotland (Kentra)	Shennan *et al.* (1995a, 1996a, 2000a)
11	57.661	−1.983	5+4	NE Scotland	Smith *et al.* (1982)
12	57.331	−1.989	4	Aberdeen	Smith *et al.* (1983)
13	56.707	−2.519	7	Montrose	Smith & Cullingford (1985)
14	56.382	−3.203	36	Tay Valley	Smith *et al.* (1985a,b); Haggart (1988a)
15	56.123	−4.153	17	Forth Valley	Robinson (1993)
16	55.870	−5.040	5+1	Ardyne	Peacock *et al.* (1978)
17	55.855	−4.494	5	Clyde	Haggart (1988a)
18	55.526	−4.678	10	Ayr	Donner (1970); Jardine (1975, 1982)
19	56.028	−2.694	3	SE Scotland	Robinson (1982)
20	55.686	−1.916	12+4	NE England (North)	Shennan *et al.* (2000d)
21	55.597	−1.728	8+2	NE England (Central)	Plater & Shennan (1992); Shennan *et al.* (2000d)
22	55.340	−1.597	26+1	NE England (South)	Plater & Shennan (1992); Shennan *et al.* (2000d)
23	54.960	−1.670	1	NE England (Tyne)	Shennan *et al.* (2000d)
24	54.629	−1.232	29	Tees	Gaunt & Tooley (1974); Tooley (1978a,b); Plater & Poolton (1992); Shennan (1992); Plater *et al.* (2000)
25	55.070	6.000	3	German Bight	Ludwig *et al.* (1981)
26	55.023	2.981	1	Dogger Bank	Shennan *et al.* (2000c)
27	54.971	−3.513	21	N Solway Firth	Jardine (1975, 1982); Lloyd *et al.* (1999)
28	54.899	−3.174	11	S Solway Firth	Huddart *et al.* (1977); Tooley (1978a); Lloyd *et al.* (1999)
29	54.395	−3.316	8	Cumbria	Huddart *et al.* (1977); Tooley (1978a)
30	54.395	−4.388	2	Isle of Man	Tooley (1977)
31	54.085	−2.958	21	Morecambe Bay	Tooley (1974, 1978a); Birks (1982); Zong & Tooley (1996)
32	53.685	−2.990	42	Lancashire	Tooley (1974, 1978a,c, 1985); Huddart (1992)
33	53.402	−3.137	15	Mersey	Tooley (1974, 1978a,c); Bedlington (1993); Cowell & Innes (1994)
34	53.300	−3.748	9	N Wales	Tooley (1978a); Kidson & Heyworth (1982); Bedlington (1993)
35	53.987	0.167	1+1	Offshore (E Of Yorkshire	Shennan *et al.* (2000c)
36	53.664	−0.065	29+4	Humber (Outer Estuary)	Gaunt & Tooley (1974); van de Noort & Ellis Long *et al.* (1998); Andrews *et al.* (2000); Metcalfe *et al.* (2000); Rees *et al.* (2000)
37	53.649	−0.661	34+30	Humber (Inner Estuary)	Smith (1958); Gaunt & Tooley (1974); van de Noort & Ellis (1995, 1997, 1998); Millett & McGrail (1987); Long *et al.* (1998); Andrews *et al.* (2000a); Metcalfe *et al.* (2000); Rees *et al.* (2000)
38	53.307	0.275	30+2	Lincolnshire Marshes	Waller (1994); Horton *et al.* (2000)
39	52.742	0.043	193+14	Fens	Shennan (1986a,b); Waller (1994); Brew *et al.* (2000)
40	52.968	0.785	31+15	Norfolk	Funnel & Pearson (1989); Andrews *et al.* (2000b)
41	52.992	1.107	1+3	Offshore (N Of Norfolk)	Shennan *et al.* (2000c)
42	53.209	2.081	2+4	Offshore (NE Of Norfolk)	Shennan *et al.* (2000c)
43	52.489	1.639	25+1	East Anglia	Coles & Funnel (1981); Brew *et al.* (1992)
44	52.473	−4.058	14	Mid Wales	Heyworth & Kidson (1982)
45	51.658	−5.066	1	S Wales (Pembrokeshire)	Heyworth & Kidson (1982)
46	51.579	−3.772	5	S Wales (Glamorgan)	Heyworth & Kidson (1982)
47	51.351	−3.025	50	Bristol Channel	Heyworth & Kidson (1982);
8	51.696	0.828	9	Essex	Greensmith & Tucker (1980); Devoy (1982); Wilkinson & Murphy (1986)
49	51.461	0.308	35	Thames	Devoy (1979, 1982)
50	51.067	1.016	34	Kent	Tooley & Switsur (1988); Long (1992); Long & Innes (1993, 1995);
51	50.838	0.329	9	Sussex	Devoy (1982); Jennings & Smyth (1987)
52	50.797	−1.349	7	Hampshire	Long & Tooley (1995)
53	50.615	−2.535	3	SW England (Dorset)	Devoy (1982)
54	50.393	−3.751	6	SW England (Devon)	Devoy (1982)
55	50.129	−5.482	6	SW England (Cornwall)	Healy (1995)

to any particular analysis. The aim is to include all radiocarbon-dated samples from palaeo-coastal environments. Many such data are rejected from further analysis because of missing information (a fault which could conceivably be corrected with further research), or because of uncertainty over the reliability of their relationship to a past sea level. In certain circumstances, samples from freshwater environments (and, using the converse argument, fully marine environments) provide important data that may be employed to test specific hypotheses because these environments must have formed inland of the palaeo-coastline and at or above the past sea level, allowing for tidal range and coastal geomorphology. The age, altitude and location of a dated freshwater sediment constrain, for example, the positions of past coastlines that constitute one of the boundary conditions for the quantitative modelling of past tidal regimes (e.g. Scott & Greenberg 1983; Gehrels *et al.* 1995; Shennan *et al.* 2000c). In reconstructing RSL history, the dated freshwater sample must lie, on a plot of altitude against age, at or above data points such as those derived from salt marsh peats whose origin was directly controlled by the palaeo-sea level. Such freshwater index points are called 'limiting dates', and their number at each of the sample sites is listed in Table 1. The same argument applies to dated samples that represent marine environments. They must lie seaward of the palaeo-coastline and below the salt marsh peats on the altitude versus age plot. Such limiting dates also contribute to achieving the research objectives of the analyses to be described in what follows.

The great majority of samples in the database for Britain come from palaeo-estuarine or coastal wetland environments where the nature of the accumulating sediment is controlled by altitude relative to the tidal regime. A second important palaeoenvironmental data source consists of the lakes or, in some cases, freshwater bogs (where the basin has filled with sediment) in coastal locations that for one or more periods between the LGM and present were marginal marine basins when the rock lip of the basin was below sea level. These marginal marine basins are frequently called 'isolation basins' since the majority record a marine phase following deglaciation of the area and a freshwater phase as sea level falls relative to the rock lip (or sill), due to isostatic rebound, thus isolating the basin from direct exposure to the sea. Fig. 4 illustrates the attributes of age, altitude and tendency for SLIs from both marginal marine basins (e to g) and palaeo-estuarine/coastal wetland environments (a to d) during a fall in relative sea level that could be due to postglacial rebound of the crust. We discuss these attributes, beginning with a fourth, location, in the subsections to follow.

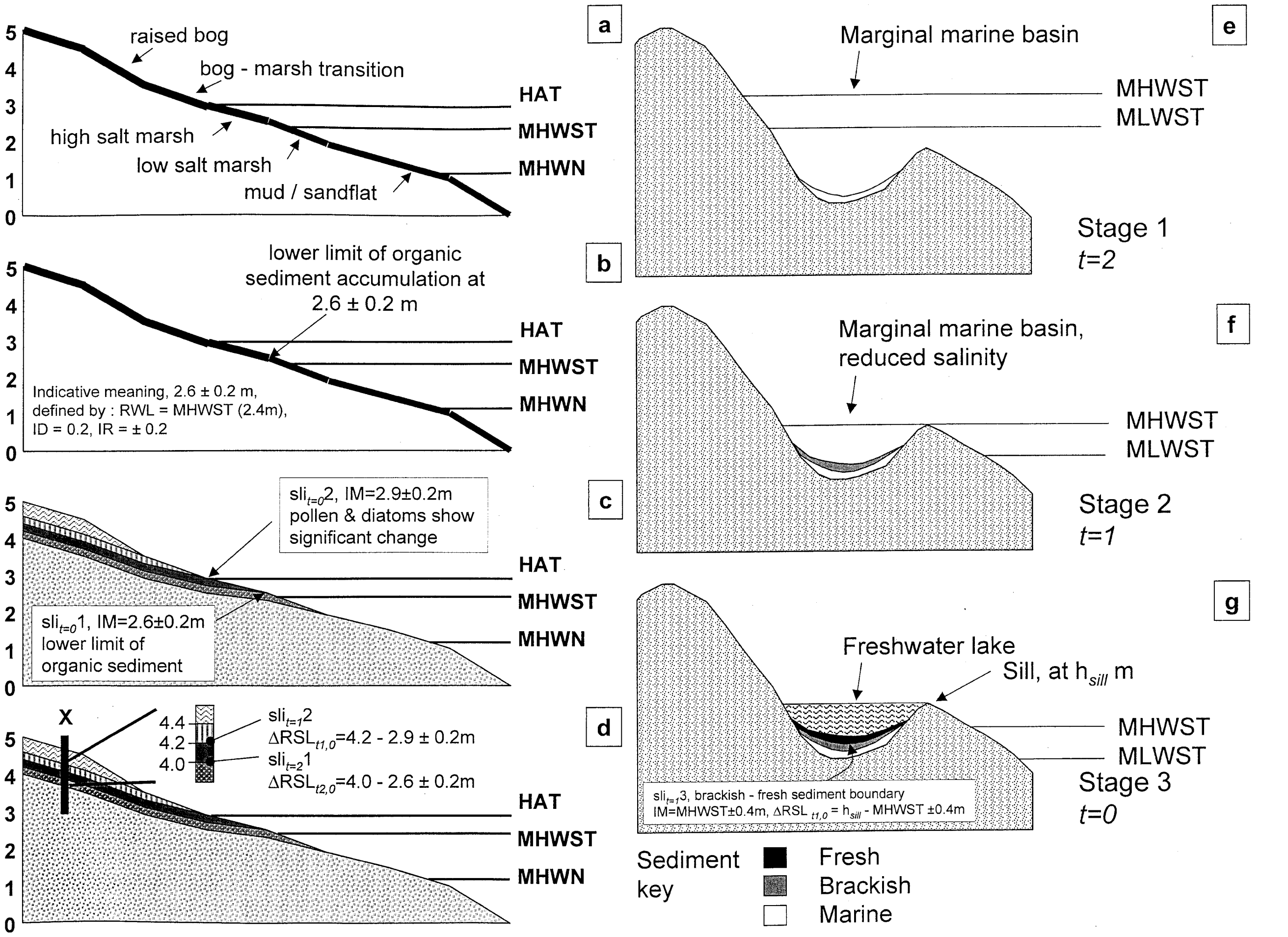


Figure 4. The sequence (a)–(d) represents the relationships that develop between various SLIs and various types of sedimentary deposit during a period of relative sea-level fall as occurs at locations in Scotland in the late Holocene due to the ongoing postglacial rebound of the crust. The sequence (e)–(g) similarly represents these relationships for a marginal basin, an 'isolation basin', which gradually becomes cut off from the sea during a similar period of falling relative sea level and is thereby transformed from a saltwater environment to a freshwater environment.

2.1 The location attribute of a sea-level index point

Once it is established that the sample has not been transported from its location of deposition, this location attribute consists simply of the geographical co-ordinates of the site from which the sample was collected. For some published data this may not be known sufficiently precisely, and where it cannot be established to within ~ 1 km the sample is rejected. For the majority of index points within the current database the location is known to 10 m or less.

2.2 The age attribute of a sea-level index point

All of the SLIs in the current database have their ages measured by radiocarbon techniques, either conventional or AMS. Archaeological artefacts and structures can also be employed to define past sea levels, but these have yet to be incorporated into the database. At present, those areas for which there exists abundant archaeological control on past sea level also possess a comprehensive record based upon radiocarbon-dated samples (e.g. Waller 1994). Other methods of dating, such as those based upon luminescence and cosmogenic isotopes (Stone *et al.* 1996; Bailiff & Tooley 2000), are still at the development stage and are frequently verified by radiocarbon-dated samples from the same or adjacent sites. The database includes both conventional radiocarbon ages and calibrated ages based on Stuiver & Reimer (1993) with the following options: calibration based on the intercepts, their method A; 95 per cent confidence limits; a laboratory multiplier of 1; the calibration data set based on bidecadal tree-ring data for 0–11 390 cal yr BP and a smooth spline interpolation of marine coral data for ages 11 400–21 950 cal yr BP. We use the calibrated ages throughout this paper which generates the LGM to the present calibration illustrated in Fig. 3.

Erosion and subsequent deposition of carbon-bearing sediment as a fraction of the dated sample results in an overestimate of age, while penetration by plant rootlets will, if not removed prior to dating, give an age that is too young. Without multiple-dated samples, sometimes on different constituents of the sediment, or other proxy measurements of age, it may be difficult to identify the effects of these contaminants (see further discussion in Strief 1979; Tooley 1978; Van de Plassche 1986. The great majority of the SLIs in the database for Britain have at least one type of corroborating evidence to support the radiocarbon age.

In many palaeo-environments, as the sea transgresses the land surface erosion of sediment may occur. The value of an SLI is limited where, for example, an unknown thickness of peat has been eroded prior to the deposition of intertidal, estuarine or marine sediment. In this case the radiocarbon (and subsequently calibrated) age of the sample from the top of the peat will be older than the sea-level-induced transgression. Unless the dated sample contains sufficient microfossils, such as diatoms, dinoflagellate cysts, foraminifera or pollen, to show that minimal erosion occurred and that it formed in a salt marsh or similar environment (directly controlled by the palaeo-sea level), it is used only as a limiting date. Where erosion occurs, the boundary between the two sediment units, for example peat and estuarine mud, is sharp, with a contact thickness that is <0.5 mm. By employing careful field techniques, which in the case of coring implies the use of equipment that provides undisturbed sediment samples, this is easily identified, and, where possible, a nearby location free of the influence of erosion should be sought. This demands a carefully constructed field sampling design with multiple sites. In some cases the field conditions or resources available limit the sampling to isolated cores, for example deep estuarine, intertidal or offshore sequences that require large and expensive drilling rigs. In some circumstances, large blocks of peat become eroded off channel sides and are then deposited in another environment. Even if there is only a single core or widely spaced cores this should be evident because both the upper and lower contacts of the peat with the adjacent sediments should show sharp boundaries.

2.3 The altitude attribute of a sea-level index point

Very few samples dated as SLIs formed exactly at the mean palaeo-sea level. Most come from environments within the upper part of the tidal range, but in total they cover the full tidal range and, for limiting dates, beyond. In order to measure RSL change, defined relative to present, it is necessary to establish the relationship of the sample to tidal range. Most analyses assume no change in tidal range since the time of deposition, but various modelling studies show that this may be an important factor, with changes ranging from decimetres (Shennan *et al.* 2000c) to metres (Gehrels *et al.* 1996) depending on the amplification of the tidal wave as the coastline shape and bathymetry change through time (as evidenced so dramatically in Fig. 2 for Britain). The relationship of a sample to the palaeo-tide level, and hence sea level, is called the 'indicative meaning' (Preuss 1979; Van de Plassche 1986). It comprises two parameters, namely the reference water level (RWL, for example mean high water of spring tides, MHWST) and the indicative range (IR: the vertical range over which the sediment could occur as it is accumulating), and the indicative difference (the vertical offset between the RWL and the midpoint of the IR). Fig. 4 illustrates these parameters. The indicative ranges of most SLIs in the British database are in the interval ± 0.2 m to ± 0.4 m (e.g. Shennan 1986b; Horton 1999).

For most of the British sites the SLIs come from palaeo-environments indicative of tidal marshes comparable to the first example in Fig. 4(a–d), with the dated sample representing accumulation of organic material close to the MHWST or above. The next most important sites, in terms of the number of index points provided, are the marginal marine basins, or isolation basins described in parts (e) to (g) of Fig. 4. SLIs from these sites can depict the whole range of the isolation (or, under RSL rise, the connection) of the basin, from the initial reduction in the full exchange of marine water when the basin is connected at all stages of the tidal cycle to the final isolation when even storm surges fail to reach into the basin (Fig. 4). Determining exactly the position of the rock sill with respect to the palaeo-tidal range represented by the dated horizon requires quantitative microfossil data. Even then, a lack of sufficient studies of present-day environments means that the indicative range is typically ± 0.4 m to ± 1.0 m (Shennan *et al.* 1999, 2000a). The higher value applies to samples for which the rate of sedimentation is low, and the vertical thickness of the sample used for radiocarbon dating represents a broad zone. The combined effects of different basin size, sill dimensions, tidal range and freshwater input upon the microfossil assemblages and hence their interpretation to give the indicative meaning of a sample remain poorly known (Shennan *et al.* 1995, 1999, 2000a).

Other factors that contribute to the height error derive from the various measurements that must be made during the sampling process (see Shennan 1986b for details). These include instrumental levelling of the site to the national datum (in Britain this is referred to as Ordnance Datum, OD). The measurement error related to this is usually ± 0.01 m for detailed surveying, but may be up to ± 0.1 m for less precise work. The precision for relating the levelling datum to local tide levels is typically ± 0.1 m, but may be as large as ± 1.5 m for offshore sites and those in large coastal lowlands along estuaries with complex tides. These figures exclude any influence of the change of tidal range through time.

The final factor that may have a significant effect on the altitude assigned to an individual SLI is sediment consolidation. Recent work by Paul & Barras 1998) advances a quantitative explanation of the impact of sediment consolidation, but their results are not yet widely applicable to late Devensian and Holocene SLIs because the necessary physical parameters of the sediments have not been measured. A few studies have included an empirical correction for sediment consolidation (e.g. Smith *et al.* 1985b), while others base their sampling design on reducing the problem by concentrating on SLIs underlain by sediments that undergo minimal or no consolidation (e.g. Jelgersma 1961, 1966; Van de Plassche 1982). In palaeo-estuarine sequences these SLIs come from thin ($\sim < 0.1$ m) salt marsh peats that overlie glacial till or other pre-Holocene sediments. These peats are frequently called 'basal peats' or 'basis peats', the latter term being applied to those that clearly formed in response to sea level in order to separate them from those basal peats that formed by local waterlogging independent of sea level. Much of the scatter in RSL graphs derives from the net effect of sediment consolidation, as illustrated in Fig. 5. Here the SLIs from basis peats all lie within the upper part of the scatter. Residuals from the general trend show statistically significant correlations with various sediment characteristics related to consolidation, such as the thickness of overburden and the thickness of sediment from immediately below the SLI to the glacial till or pre-Holocene (Shennan *et al.* 2000b). For locations where the SLIs derive from such palaeo-estuarine environments, best-fit RSL predictions should lie towards the top of the scatter (e.g. as for the Fenlands data set shown in Fig. 5). Sediment consolidation has no influence on SLIs from marginal marine basins because the rock sill is the important height, not the height of the sediment within the basin (Fig. 4).

The total height error, E_h, illustrated in the figures to be presented in Section 4 of this paper is calculated on the basis of the expression

$$E_h = (e_1^2 + e_2^2 \ldots + e_n^2)^{1/2}, \qquad (5)$$

where $e_1 \ldots e_n$ are the individual sources of error discussed above.

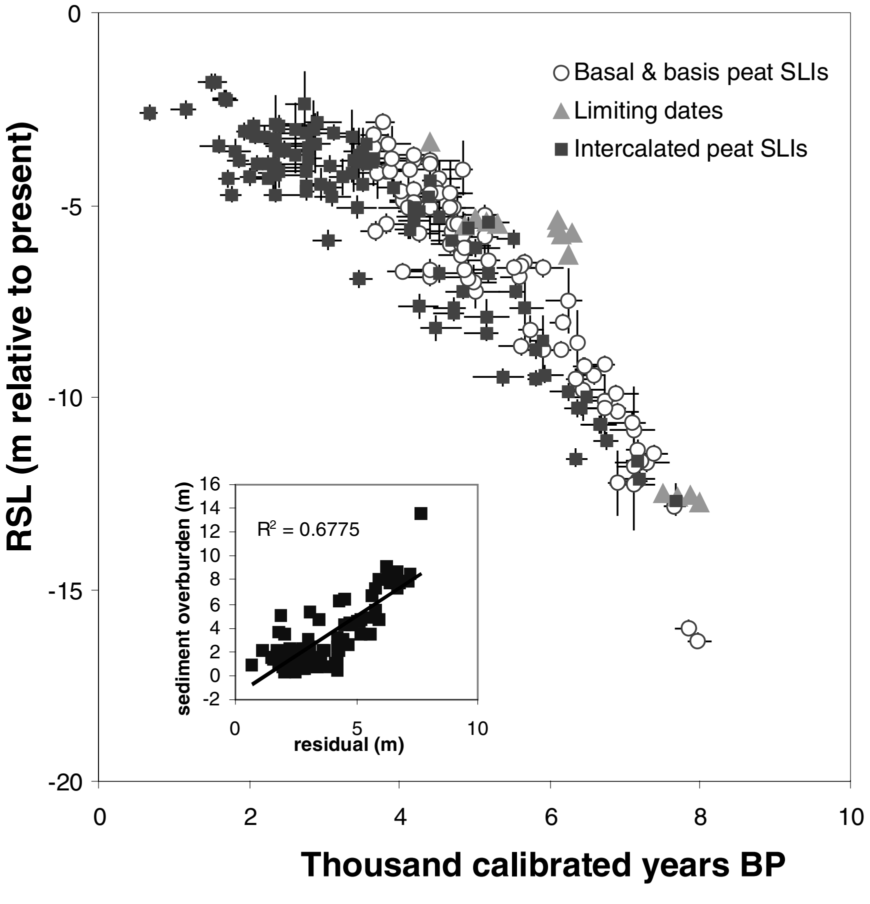


Figure 5. Complete collection of data from the Fenlands site (no. 39 on Fig. 2) illustrating the relative levels in the sedimentary sequence of basal and basis peats (defined in the text), of limiting dates (defined in the text), and of SLIs associated with intercalated peats.

2.4 The tendency attribute of a sea-level index point

The tendency of an SLI describes the increase (positive sea-level tendency) or decrease (negative sea-level tendency) in marine influence recorded by the SLI. Thus, all of the example SLIs from the raised tidal marsh sequence and the isolation basin depicted in Fig. 4 are characterized by negative sea-level tendencies. The converse situations, a transgression of a salt marsh sequence by tidal flat deposits or the ingress of marine water into an isolation basin, represent positive sea-level tendencies. Most analyses of the pattern of sea-level tendencies, either within (e.g. Shennan 1986b; Haggart 1988a) or between estuaries (Shennan *et al.* 1983b), focus on the relationships between horizontal shifts in the coastal sedimentary environments, sediment accumulation, consolidation, and the potential driving mechanism of relatively small, ~ decimetres, vertical changes in sea level. Allen (1990a,b,c,d, 1995, 1997) describes a quantitative model to represent these important relationships but they have yet to be applied widely to data other than those used, from the Severn estuary, to develop the model.

Beyond these estuary-scale considerations, the major applications of identifying sea-level tendencies in testing RSL model predictions derived from quantitative models of the glacial isostatic adjustment process are twofold. First, observations of a change in sign of the tendency are very useful in determining the turning points in the non-monotonic RSL curves for areas such as Scotland, where such non-monotonicity derives from the competition between the crustal rebound induced by the removal of ice and the rise of relative sea level due to the melting of ice from distant locations. For example, a detailed foraminifera record from Loch and Corr in NW Scotland (Lloyd *et al.* 1999) dates the switch from negative to positive sea-level tendency in the early Holocene. Even though there is a wide error band for the altitude of RSL, because of the lack of modern analogues for the depth of seawater represented by the fossil assemblage, the trend to shallower water is unequivocal. For RSL predictions to conform with these data they should be showing the start of a rise in RSL at that time, as well as falling within the vertical error band (Shennan *et al.* 2000a). Second, sea-level tendencies show whether the RSL was slightly above or below the present level in the last few thousand years, thus identifying the geographical limits of the transition from relative land uplift to relative submergence, as has been illustrated for a transect along the east coast of England (Shennan *et al.* 2000b). The current database comprises a total of 934 sea-level index points and 91 limiting dates, as summarized in Table 2.

3 THE GLOBAL GEOPHYSICAL THEORY OF THE GLACIAL ISOSTATIC ADJUSTMENT PROCESS

The model that we will employ as the basis for interpretation of the relative sea-level data from sites in the British Isles is that which has been under continuous development in Toronto since the initial publication of Peltier (1974). In that paper, a detailed mathematical formalism was developed that could be employed to compute the impulse response of a radially stratified and linearly viscoelastic Maxwell model of the Earth to excitation by an applied surface mass load. Subsequent further developments of this theory (Peltier 1976; Peltier & Andrews 1976; Farrell & Clark 1976) culminated in the formulation of what has come to be called the 'sea-level equation' (SLE), an integral equation whose solution delivers direct predictions of the time-dependent separation between the mean level of the surface of the sea (the geoid of classical geodesy) and the surface of the solid Earth. The most primitive form of this integral equation for RSL history can be written as

$$S(\theta,\lambda,t) = C(\theta,\lambda,t)\left[\int_{-\infty}^{t} dt' \iint_{\Omega} d\Omega' L(\theta',\lambda',t') \times \left\{\frac{\phi^{\mathrm{L}}}{g}(\mathbf{r}-\mathbf{r}',t-t') - R^{\mathrm{L}}(\mathbf{r}-\mathbf{r}',t-t')\right\} + \frac{\Delta\Phi(t)}{g}\right], \tag{1}$$

in which $C(\theta, \lambda, t)$ is the so-called ocean function which equals unity at all points on the Earth's surface that are 'in' the global ocean and zero otherwise. The space–time functions ϕ^{L} and R^{L} are viscoelastic Green's functions, for the perturbation to the gravitational potential which defines the geoid and the radial displacement perturbation, respectively, for the surface loading problem whose static elastic forms are identical to those discussed by Farrell (1972). The function $L(\theta, \lambda, t)$ is the longitude-, latitude- and time-dependent surface mass load per unit area, which has the composite form

$$L(\theta,\lambda,t) = \rho_{\mathrm{I}} I(\theta,\lambda,t) + \rho_{\mathrm{w}} S(\theta,\lambda,t)\,, \tag{2}$$

in which ρ_{I} is ice density and I is the ice thickness, and ρ_{w} is the water density and S is the perturbed 'water thickness' measured with respect to the (deforming) sea floor. Because of the composite form of (2) it will be clear that eq. (1) is an integral equation for the relative sea-level history S, this unknown field appearing not only on the left-hand side but also as an argument of the triple convolution integral on the right-hand side. The last term on the right-hand side of (1), namely $\Delta\Phi(t)/g$, is an exactly calculable time-dependent correction that is required to ensure that mass is conserved during the processes of surface load redistribution and Earth deformation. Peltier (1998a) has provided a recent review of the derivation of the mathematical form of this term.

As discussed in some detail in Peltier (1998a, 1999), the form (1) of the sea-level equation may be modified slightly to incorporate the influence on sea level of the changing rotational state of the planet. This modified sea-level equation takes the form

$$S(\theta,\lambda,t) = C(\theta,\lambda,t)\left[\int_{-\infty}^{t} dt' \iint_{\Omega} d\Omega' \{L(\theta',\lambda',t')G^{\mathrm{L}}(\mathbf{r}-\mathbf{r}',t-t') + T(\theta',\lambda',t')G^{\mathrm{T}}(\mathbf{r}-\mathbf{r}',t-t')\} + \frac{\Delta\Phi(t)}{g}\right], \tag{3}$$

in which $T(\theta', \lambda', t')$ is the variation of the centrifugal potential caused by the changing rotational state of the planet, and $G^{\mathrm{T}}(\mathbf{r}-\mathbf{r}', t-t')$ is the so-called tidal loading Green's function (see Dahlen 1996 for a discussion of this additional source of sea-level excitation). The Green's functions G^{L} and G^{T} have infinite series expansions in terms of the usual surface load and tidal Love numbers as

$$G^{\mathrm{L}}(\gamma, t) = \frac{ag}{m_{\mathrm{e}}} \sum_{\ell=0}^{\infty} [1 + k_\ell^{\mathrm{L}}(t) - h_\ell^{\mathrm{L}}(t)] P_\ell(\cos\gamma)\,, \tag{4a}$$

$$G^{\mathrm{T}}(\gamma, t) = \frac{1}{g} \sum_{\ell=0}^{\infty} [1 + k_\ell^{\mathrm{T}}(t) - h_\ell^{\mathrm{T}}(t)] P_\ell(\cos\gamma)\,, \tag{4b}$$

in which a is the Earth's mean radius, g is the surface gravitational acceleration, m_{e} is Earth's mass, P_ℓ are the usual Legendre polynomials, and (k^{L}, h^{L}), (k^{T}, h^{T}) are, respectively, the time-dependent surface load and tidal Love numbers (see Farrell 1972 for a discussion of the equivalent elastic forms). Each of these four Love numbers has a time dependence that can usually be written in the form of a finite sum of purely exponential relaxations as (Peltier 1974), for example,

$$h_\ell^{\mathrm{L}}(t) = \sum_{j=1}^{J} r_j^\ell \exp(-s_j^\ell t)\,. \tag{5}$$

The s_j^ℓ may be computed, following Peltier (1976), as the zeros of a 'secular function' in the domain of the Laplace transform variable s, and these zeros define a set of 'normal-mode poles' that is in common to each of the four Love numbers. The r_j^ℓ however, are the residues at these poles (the normal-mode amplitudes) and are computed for each of the Love numbers using the calculus of residues methodology described in detail in Peltier (1985). This methodology for computation of the modal amplitudes [which was recorded in Wu (1978) but not applied therein as it was at first found to deliver unstable results] was first implemented correctly for the complete numerical model in the Peltier (1985) publication. The complete viscoelastic kernel G^{L} for the sea-level equation was first computed in Peltier & Andrews (1976) using a more direct but less sophisticated collocation method to obtain the time dependence of the Love numbers from their Laplace transforms. In circumstances in which the radial structure of the model becomes sufficiently complex, current practice is to employ a hybrid methodology in which the amplitude of most modes is determined on the basis of the calculus of residues methodology but in which other *ad hoc* poles may be introduced in such a way as to ensure that the inverse Laplace transforms are acceptably accurate. No purpose will be served by further reviewing the details of the mathematical methods that have been developed to solve (3). Suffice it to say that the spherical-harmonics-based pseudo-spectral methodology that is currently employed is a further refined version, described in Peltier (1998a, 1999), of that discussed in Mitrovica & Peltier (1991). The refinement involves the use of an iterative method described in Peltier (1994) to simultaneously solve for the evolving coastline, represented by the boundary of $C(\theta, \lambda, t)$, and for the influence of the changing rotation, as well as for the global history of relative sea-level change $S(\theta, \lambda, t)$ itself.

The primary assumption on which this global viscoelastic theory of the GIA process is based is that the internal structure of the Earth can be adequately approximated as being radially stratified. From the very earliest efforts in the development of this theory (Peltier 1974), it has been a central goal of the analyses performed with it to attempt to discover whether there might not be revealed thereby a number of systematic misfits of the theory to the observations—misfits that one might be obliged to ascribe to the influence of lateral heterogeneity of viscosity.

One such possible systematic error, pointed out by Tushingham & Peltier (1991), concerned a significant over-prediction of the amplitude of glacial forebulge collapse as this is recorded by a dense set of ^{14}C-dated RSL histories from sites located along the east coast of the continental United States. The initial theoretical predictions that misfit the US east coast observations in this way were those that had been made using the early ICE-3G (VM1) model of Tushingham & Peltier (1991). Analyses thereafter appeared in the literature in which various authors sought to show that this systematic error could conceivably be due to the influence of lateral viscosity heterogeneity. The originally identified misfits in this region were, however, associated with predictions made with the extremely simple model of the radial variation of viscosity that has since come to be denoted VM1 for Viscosity Model 1. As demonstrated in Peltier (1996a) and Peltier & Jiang (1996, 1997), when this simple model of the radial viscoelastic structure was replaced by a more complex model labelled VM2 that was determined on the basis of formal Bayesian inversion of a subset of the GIA information that did not include these US east coast data, the previously identified misfits of the radially stratified model VM1 were very nearly entirely eliminated. The primary difference between VM1 and the revised model that was denoted VM2 lay in the fact that the upper mantle and transition zone viscosity in the latter was reduced to a value near 0.4×10^{21} Pa s from its 1×10^{21} Pa s value in VM1. At present we are therefore in a very strong position to argue that no systematic errors have yet been identified that argue unambiguously for the importance of lateral viscosity heterogeneity of the mantle in properly reconciling the totality of the available data related to the process of GIA.

That no such unambiguous evidence currently exists cannot of course be construed to suggest that the viscosity of the mantle varies only with radius. If the process of plate tectonics is indeed controlled by thermal convection in the mantle, the possibility that the viscoelastic structure of the Earth is spherically symmetric must be rejected on *a priori* grounds. Since convective mixing is driven by lateral variations of temperature, and since the creep of a polycrystalline solid like the Earth's mantle is a thermally activated process, it is clear that strong lateral variations of viscosity must exist in the system. It may well be the case, however, that the horizontal length-scales over which significant viscosity variations occur are relatively small, say of the order of the lithospheric thickness and the width of the down-going slabs observed in seismic Benioff zones, in which case the GIA process, which occurs on relatively large horizontal scales at locations remote from subduction zones and mid-ocean ridges, would primarily interrogate the adiabatic and thus radially stratified cores of the convection cells themselves. Evidence that could be construed to oppose this view comes from models of 3-D mantle structure that have recently been delivered by seismic tomography. If one were to accept the models of Woodhouse & Dziewonski (1984) and Dziewonski (1984 and subsequently) as definitive in this regard, it would follow from the observed dominance of large spatial-scale lateral heterogeneity in these models that viscosity should also vary on the same spatial scales. On the other hand, the higher-spatial-resolution models of Van der Hilst *et al.* 1997) and others reveal dominant structures on much smaller spatial scales. Since it seems clear that the models generated by Dziewonski and co-workers are, to good approximation, more accurate at low wavenumbers but nevertheless low-pass-filtered versions of their higher-resolution counterparts, it seems entirely plausible that the influence of strong lateral heterogeneity of viscosity may be very difficult to detect in glacial isostatic adjustment observations, precisely because it is confined to regions of relatively small horizontal scale.

Our intention herein is therefore to continue to subject the spherically symmetric model ICE-4G (VM2) to further intensive testing, using the British Isles data set to provide the basis of such further analysis. Just as was the case with the data set from the US east coast, none of the data from the UK were employed as the basis for construction of the VM2 viscosity model. They will, however, be employed to further refine it in what follows, as we shall see.

4 ANALYSIS STRATEGY AND THE RESULTS OF DATA–MODEL COMPARISON

Since the procedure whereby the previously discussed RSL observations are to be compared with the predictions of the theoretical model differs significantly from that which has typically been adopted in the series of papers by Lambeck and co-workers in previous analyses of RSL data for the British Isles, and since these differences will be seen to be important, it will be useful prior to the presentation of the results to enumerate these differences of approach. First, as discussed above, the global viscoelastic theory for postglacial RSL history is based upon the assumption that the parameters of the viscoelastic model, consisting of density as well as the elastic Lamé parameters and viscosity, are functions of radius only. Because it is an important question whether a model of this kind will enable a viable explanation of the majority of the data associated with the global GIA process to be constructed, we will therefore insist in what follows on testing the impact that any modification to the radial structure of the VM2 model (that may be suggested by the UK data set) would have upon our ability to reconcile observations from other locations.

Second, we perform the model–data comparisons on a calendar-year-based timescale rather than on the ^{14}C timescale that was adopted in the recent paper by Shennan *et al.* (2000a) and in all previous work by Lambeck and colleagues. Although it has been suggested (Lambeck 1998) that viscosity estimates on the ^{14}C timescale might be adjusted to the physical viscosity through an *a posteriori* constant multiplicative correction and the attribution of suitable error terms, we argue herein that this approach is not justifiable since the observations of Scottish rebound that we intend to interpret depend as strongly upon lithospheric thickness as they do upon viscosity, so that the use of the ^{14}C timescale would introduce significant distortion of the interpretation of observations that extend to sufficiently great time into the past. We will therefore present the observations on the calendar-year timescale by applying the CALIB 3.0 procedure of Stuiver & Reimer (1993) as discussed in Section 2 of this paper. In order to provide an explicit example, independent of the current data set, of the possible importance of this conversion, we show in Fig. 6 a comparison of relaxation times deduced for the large set of ^{14}C-dated relative sea-level histories from the Laurentian platform recently discussed by Dyke & Peltier (2000), for which purpose the relaxation times are determined by application of a Monte Carlo method to obtain the best-fitting parameters of an exponential model of the relaxation process. Whereas in Dyke and Peltier the relaxation times were determined on the basis of data on the ^{14}C timescale, this Figure compares the relaxation times inferred in this way with those obtained when the individual sample ages in each time-series are first translated to calendar years. All of the RSL curves from this geographical region are strikingly exponential in character, and the observed relaxation times are directly determined by the model mantle viscosity over the range of depths to which the data from the site in question are most sensitive (for the Canadian data set this range consists primarily of the upper $\sim$700 km of the lower mantle, as explicitly discussed in Peltier 1996a). The relaxation times on the ordinate labelled Tsp are the relaxation times determined for each of the time-series after the individual ^{14}C-dated observations have been calibrated into calendar years using the CALIB 3.0 procedure. The abscissa labelled Tcp, on the other hand, shows the relaxation times determined from the same data sets when the ages of the individual measurements are kept on the ^{14}C timescale. This comparison clearly demonstrates that the magnitude of the correction that one would be obliged to make to the model mantle viscosity (say based upon the quadratic fit with parameters in the Figure legend), if one were to measure the relaxation time on the ^{14}C timescale, is a function of the inferred relaxation time. If the inferred relaxation time on the ^{14}C timescale is short then the correction is small. However, when the inferred relaxation time is large then so is the correction, the sense of which is such that the 'correct' relaxation time is longer. If the data were not corrected in this way we would therefore infer values of mantle viscosity that were biased to lower values than actually exist. In the case of the Scottish rebound data to be analysed herein the situation is very much more complicated, as the observations are as strongly influenced by lithospheric thickness as by mantle viscosity, and as strongly influenced by the rate and amount of RSL rise due to distant sources of meltwater as by lithospheric thickness. The procedure we will follow is therefore to first transform the observations from ^{14}C time to the calendar-year timescale, and then to perform the required comparison of the observations with the theory, the latter clearly delivering results on uniform 'Newtonian time', which is also measured in calendar years.

The third, and perhaps most important, way in which our analyses of the UK RSL data will differ from those previously performed is that we explicitly recognize the extreme sensitivity of the RSL time-series from this region to the details of the deglaciation process at other locations on the Earth's surface. As we will show, it is not possible to infer a local radial viscoelastic structure of the planet for this geographical region except in conjunction with an explicit global model of the deglaciation of the planet as a whole. No detailed demonstration of this fact has previously appeared in the literature on the interpretation of UK RSL data on the basis of the predictions of the global viscoelastic model of the isostatic adjustment process. Our explicit demonstration of the importance of these influences will constitute one of the more original contributions of this paper.

4.1 The impacts of shallow viscoelastic structure and ice-sheet thickness on RSL histories from the British Isles

As a first step in the investigation of the impact of the constraint upon ice-sheet thickness, lithospheric thickness and sublithospheric viscosity provided by the British Isles data, we

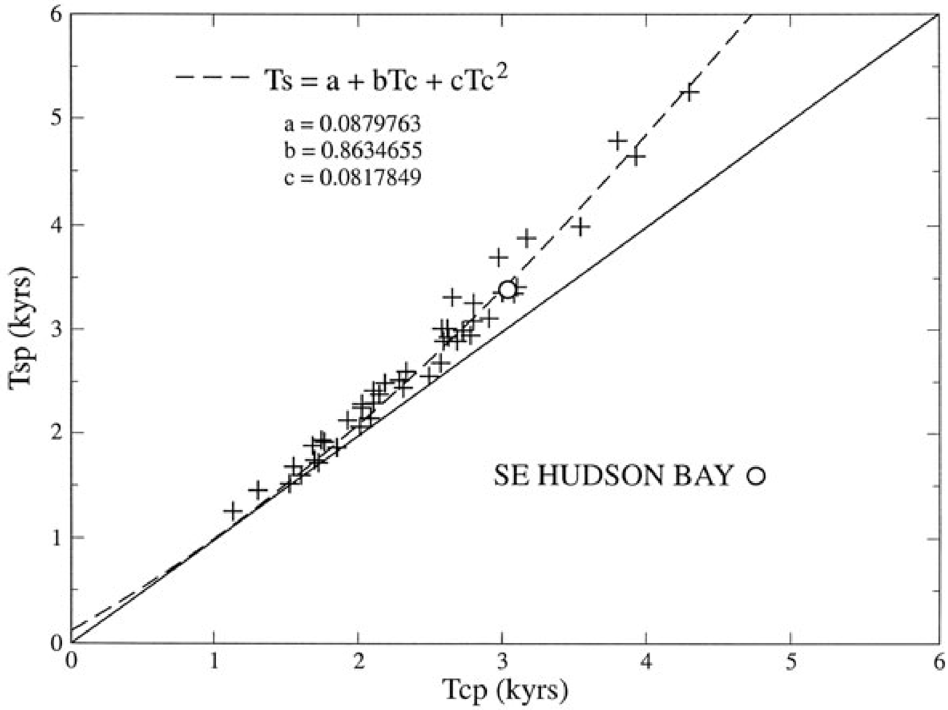


Figure 6. Comparison of exponential relaxation times inferred for a large set of ^{14}C-dated relative sea-level curves from the Laurentide platform previously discussed by Dyke & Peltier (2000). On the ordinate the relaxation times for the data from individual sites are shown in sidereal years. To obtain these results, each of the ^{14}C dates in an individual time-series was first transformed to sidereal age using the Stuiver & Reimer (1993) calibration, and the relaxation time was inferred by employing a Monte Carlo procedure to fit a simple exponential model to the data. The value of the abscissa co-ordinate of each point was obtained by directly fitting the exponential model to the individual time-series on the ^{14}C timescale. Clearly, the relaxation times on the ^{14}C timescale are shorter than those on the sidereal timescale, and the difference increases rapidly with increasing relaxation time. The relaxation time of 3.4 kyr for the complete SE Hudson Bay data set obtained by Peltier (1998a) is shown as the open circle. This RSL curve is shown in Fig. 17 together with the fit to it of the ICE-4G (VM2) prediction.

show in Fig. 7(a) the quality of the fits to the Fennoscandian relaxation spectrum originally inferred by McConnell (1968) of a series of models that differ only in terms of their shallow viscoelastic structures. Wieczerkowski *et al.* (1999) have recently confirmed the essential validity of McConnell's inference of the dependence of relaxation time upon the spherical harmonic degree (horizontal wavenumber) of the surface deformation at long wavelengths, although they argue that the disagreement between the 5-beach and 6-beach fits shown in Fig. 7(a) underestimates the magnitude of the error that could be associated with individual wavenumber estimates, and, furthermore, that the relaxation-time estimates at highest spherical harmonic degree are entirely unreliable, as previously noted. Since the VM2 viscosity model, whose relaxation spectrum is shown in Fig. 7(a), was designed so as to fit the long-wavelength components of this spectrum reasonably well, and because this model was thereafter shown to eliminate the significant misfits to US east coast RSL data that had previously been shown to be delivered by the VM1 viscoelastic model, we will insist that any modification to the shallow structure of VM2 must be such that the US east coast data continue to be fit by the new model.

In Fig. 7(a) we show relaxation spectra for six models. The three closely spaced spectra are for the original VM2 model and for two others which differ from VM2 in both lithospheric thickness and sublithospheric viscosity. These models with $L = 90$ km and $L = 100$ km have had their sublithospheric viscosity modified from VM2, so that they have Fennoscandian relaxation spectra essentially identical to that of VM2 itself. The viscoelastic structures of these models are shown in Fig. 7(b). It will be clear by inspection of this Figure that, in order to compensate for the impact of thinning the lithosphere, the immediately sublithospheric viscosity must be reduced from the elevated level characteristic of VM2 to recover the same fit to the relaxation spectrum.

Also shown in Fig. 7(a) are the spectra for model VM1 and for a further model denoted VM2 L90, in which the relatively high viscosity immediately beneath the lithosphere that is characteristic of VM2 is maintained and simply extended to the shallower depth of the lower boundary of the 90 km thick lithosphere that obtains in this case. Inspection of Fig. 7(a) will show that the model with lithospheric thickness $L = 90$ km and VM2 sublithospheric viscosity structure has a Fennoscandian relaxation spectrum which matches that of VM2 at long wavelengths and that of VM1 at short wavelengths. As will become clear in what follows, it is this model that will allow us to fit the sea-level data using a representation of the Scottish ice-sheet that reconciles the Ballantyne *et al.* (1998) constraint on maximum ice thickness and possesses VM2 sublithospheric viscosity structure. Since Fennoscandia and the British Isles are so close in geographical position, it is clearly expected that the viscoelastic models for the two regions should be similar.

Fig. 8 compares the two models of the LGM distribution of glacial ice over the British Isles that will prove to be critical to the ensuing discussion. The ice-thickness distribution shown in

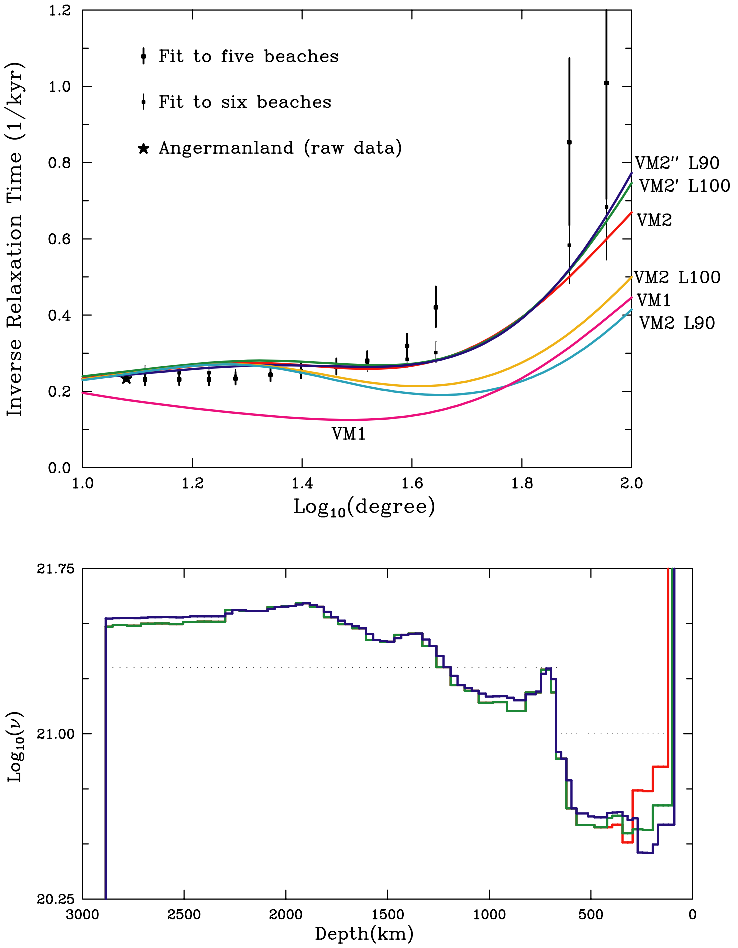


Figure 7. (a) The relaxation spectrum for Fennoscandian rebound according to the 5-beach and 6-beach estimates of McConnell (1968) showing the apparent relaxation time as a function of the spherical harmonic degree of individual spatial scales in the deformation spectrum. Also shown, by the star symbol, is the relaxation time (in sidereal years) inferred on the basis of the relative sea-level history at Angermanland, Sweden. Theoretical predictions of the Fennoscandian relaxation spectrum are shown for the sequence of radial viscosity models shown in (b). The primary models employed for the detailed RSL calculations discussed in the text consist of VM2, which has a lithospheric thickness of 120 km, and VM2 with $L = 90$ km.

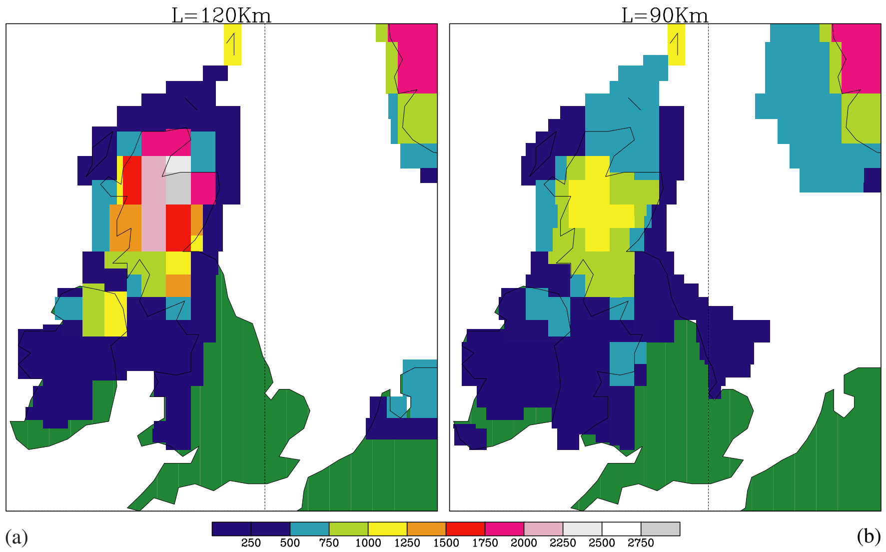


Figure 8. (a) LGM ice thickness over Great Britain in the ICE-4G model of the LGM to Holocene deglaciation event. (b) The thickness distribution for the modified model constructed so as to satisfy the constraint on the maximum thickness of Scottish ice described by Ballantyne *et al.* (1998).

Fig. 8(a) is that employed in the ICE-4G model of deglaciation. This model has a maximum thickness over the Scottish Highlands that is somewhat thicker than 2 km so that this ice-sheet is very close in this property to the early glaciological reconstruction of Boulton *et al.* (1977). In the Boulton *et al.* (1977) model, however, the ice-sheet was assumed to extend across the North Sea and to join with the Fennoscandian ice-sheet on its western flank. For the second of the ice-distribution models shown in Fig. 8(b), the thickness over the Scottish Highlands is sharply reduced by a factor of approximately 2 such that the maximum thickness of ~ 1 km is in much better accord with the recently published *a priori* constraint of Ballantyne *et al.* (1998). This model is now extremely similar to the further glaciological reconstruction of Boulton *et al.* 1985), which was based on the assumption of very mobile conditions at the base of the ice mass and which did not have the ice-sheet extending to the east so as to merge with the Fennoscandian complex. In the case of both of the LGM ice distributions shown in Fig. 8, the retreat isochrons were constrained to those of Andersen (1981), although the model shown in Fig. 8(a) does not include the well-constrained extension to the south of Scotland along the east coast of England. Of interest in the remainder of this paper is the issue of whether the ice distribution in Fig. 8(b), when coupled with a radial viscoelastic structure that consists of the VM2 viscosity distribution but a lithospheric thickness of $L=90$ km, will deliver good fits to the RSL observations. Since $L=90$ km is in accord with the inference of DeLaughter *et al.* (1999), who obtained $L=95\pm10$ km, if we are able to demonstrate that consistent results can be obtained using these two different methodologies, this will suggest that the much reduced lithospheric thickness of $L=65$ km preferred by Lambeck and colleagues is open to question.

Further insight into the nature of these alternative ice-sheet reconstructions is provided in Fig. 9(a), which shows the bulk melting histories for these two models of British Isles deglaciation in the form of the eustatic (globally averaged) sea-level histories produced by them. Of particular note is that the contribution of these ice-sheets to the approximately 120 m rise of global sea level that occurred during the most recent glacial–interglacial transition is extremely slight. Even the ice complex in the thick-ice reconstruction of ICE-4G delivers a global sea-level rise that is little more than 0.5 m. The eustatic sea-level rise induced by the melting of the thinner complex that fits the Ballantyne *et al.* constraint reasonably well is 20 per cent less, being only 0.40 m. The eustatic sea-level rise induced by the complete set of components that constitute the ICE-4G model is shown in Fig. 9(b), together with the individual contributions from the various geographical regions. It should be noted that this Figure shows only the contributions to global sea-level rise due to the 'explicit' components of the ICE-4G model, the 'implicit' contributions, discussed in Peltier (1998a,b,c), have no significant impact on the relative sea-level prediction and are not shown here. They are required, however, in order to understand why the 'explicit' ice amounts that are employed in solving the sea-level equation do not correspond to the approximately 120 m of eustatic sea-level rise that actually characterized the LGM to Holocene deglaciation event.

Also evident by inspection of Fig. 9(a) is the slight drop in sea level that was caused by the re-advance of the Scottish ice-sheet that occurred during Younger Dryas time, 12 000 calendar years ago, and which is locally referred to as the Loch Lomond Stade. The outline of this region of renewed glaciation is shown in Fig. 1 of Ballantyne *et al.* (1998) and is essentially perfectly coincident with the core of the ice mass in the two LGM ice-sheet reconstructions shown in Fig. 8, as one would expect since this is the region of highest topographic elevation. Notable also in Fig. 9(a) is the fact that the re-advance in ICE-4G was assumed to have occurred significantly later than Younger-Dryas time and to have involved a significantly greater degree of re-glaciation than in the second and thinner reconstruction that we have produced in order to satisfy the Ballantyne *et al.* (1998) thickness and age constraints. The detailed characteristics

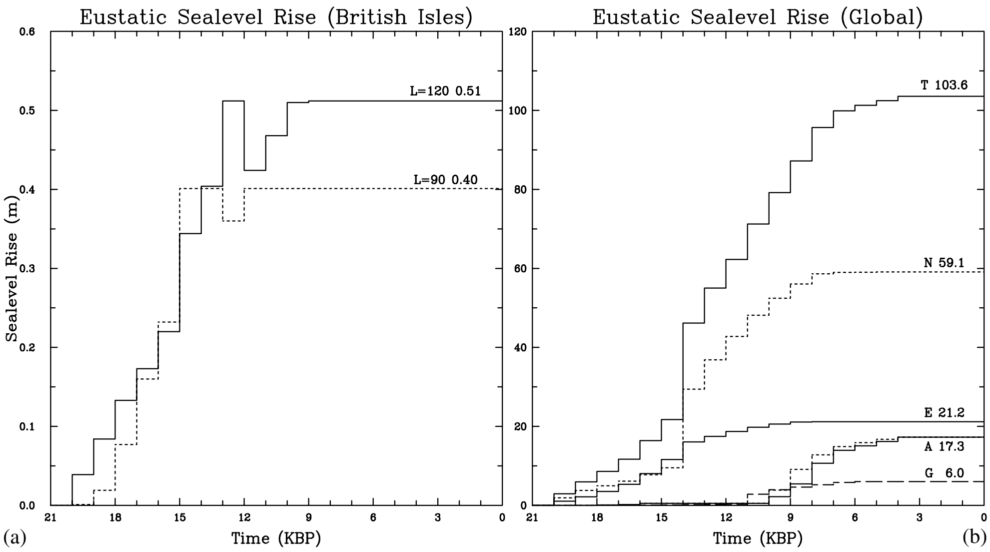


Figure 9. (a) Eustatic sea-level histories delivered by the melting of each of the models of the UK LGM ice-sheets shown in Fig. 8. Note that neither of these melting events causes more than ~ 0.5 m of global sea-level rise, and that each includes a slight (Loch Lomond stade) re-advance during the Younger Dryas period. (b) Total eustatic sea-level history induced by the ICE-4G model of the last deglaciation event of the late Pleistocene ice-age. Also shown are the contributions to the global eustatic history from each of the four primary locations at which large accumulations of LGM ice existed (N: North America, E: European, G: Greenland, and A: Antarctica and Patagonia). Note also the modified version of the Antarctic history shown as the dashed line.

of the timing and extent of the Loch Lomond re-advance and the preceding ice-melt history do have a significant influence on the results to be presented in what follows. Since our analyses are being performed with a temporal discretization characterized by a time-step of 1 kyr, however, there is a limit to the extent to which we are able to explore the impact on the inference of the radial viscoelastic structure of subtle variations in the features of the re-advance.

Direct comparisons of model predictions of relative sea-level history and the observations for virtually every site in the new, extensive, post-LOIS data base for Great Britain are shown in Fig. 10 for three different models. These three models differ in the combination of radial viscoelastic structure and local deglaciation history employed as input to the solution of the sea-level equation, but all solutions have been computed employing a high-resolution truncation of the spherical harmonic representations of the model fields to degree and order 512. The long-dashed line in each of the plates of this Figure denotes the prediction for the original ICE-4G (VM2) model. For present purposes the original ICE-4G model has in fact been modified slightly by delaying the Antarctic Ice Sheet component of the melting history by 1 kyr and slightly reducing the eustatic sea-level rise delivered by this region to ~ 17 m from the ~ 22 m that characterized ICE-4G (see Fig. 37 of Peltier 1998a for a detailed disaggregation of the ICE-4G eustatic curve). On each of the site-specific comparisons of observations and theoretical predictions shown in Fig. 10, the dotted lines denote the predictions based on the use of the thin British ice-sheet version of the ICE-4G model, to which we will subsequently refer as ICE-4GUK, but which employs the same VM2 model of the viscoelastic structure which has a lithospheric thickness of 120.7 km. The solid lines show the predictions of RSL history delivered by the glaciation history ICE-4GUK in conjunction with VM2 sublithospheric viscosity but a lithospheric thickness of $L=90$ km.

Inspection of this set of comparisons demonstrates very clearly that the latter model is preferred at most of the locations (in the north) at which the predictions of the model are sensitive to the variations introduced. Everywhere south of approximately site 31 in Fig. 2, however, the sensitivity of model predictions to these variations is extremely slight and all of the models are seen to fit the observations extremely well. At the northern locations the best fits to the data are seen to be delivered either by the original ICE-4G (VM2) model or by the model with both reduced lithospheric thickness and reduced ice thickness. The data from the Arisaig site, however, which are based upon extremely accurate isolation basin analyses and which extend over the longest period of time, are quite unambiguous in their strong preference for the latter model. The data sets from the Forth Valley, from the Tay Valley, and from Ardyne, at which the strength of the non-monotonicity in the RSL record is most strongly developed, echo this preference. To the extent that the thick ice–thick lithosphere ICE-4G (VM2) model is ruled out by the Ballantyne *et al.* constraint on ice thickness, however, only the thin ice–thin lithosphere solution is acceptable. A better appreciation of the extent to

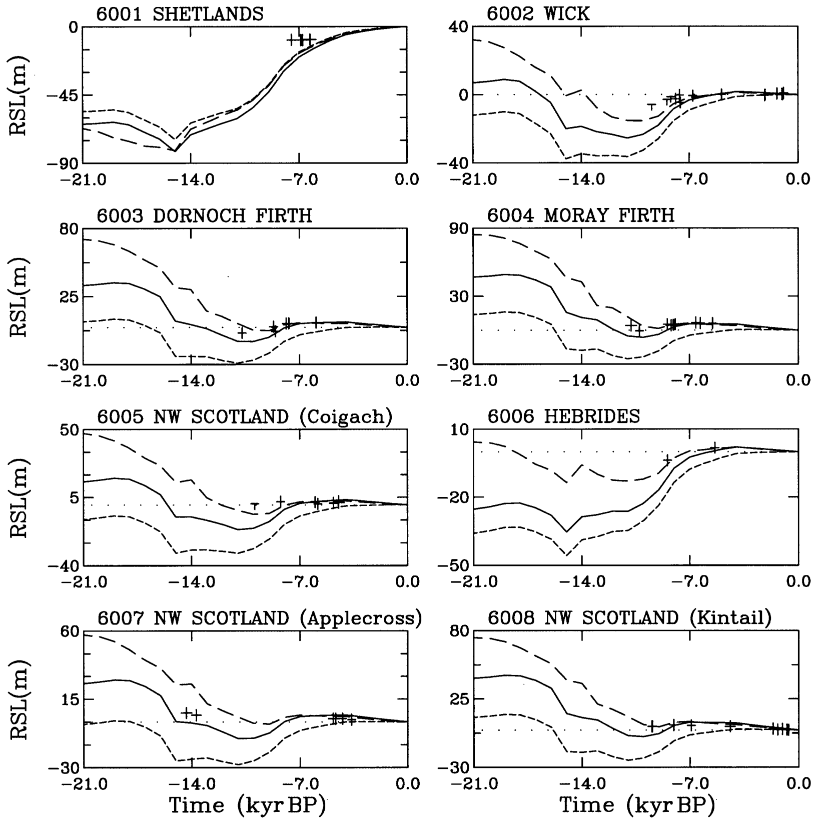


Figure 10. Complete sequence of comparisons of observed relative sea-level histories from UK locations with theoretical predictions for each of the three combinations of shallow viscoelastic structure and local deglaciation models discussed in the text. The solid curves are the predictions for the model with thin ice and thin lithosphere. The long-dashed lines are for the original ICE-4G (VM2) model, and the short-dashed lines are for the thin-ice model with the original VM2 ($L=120.7$ km) viscoelastic structure.

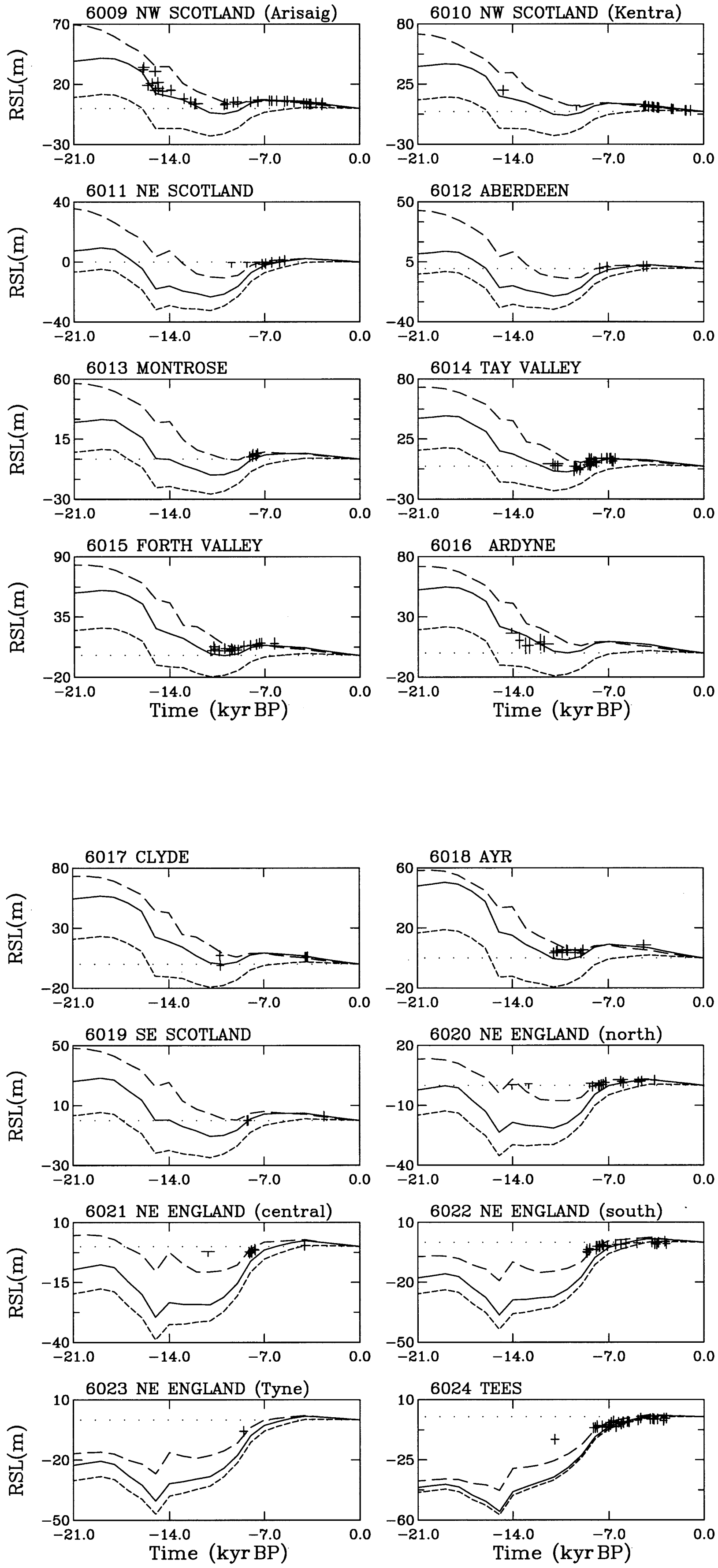
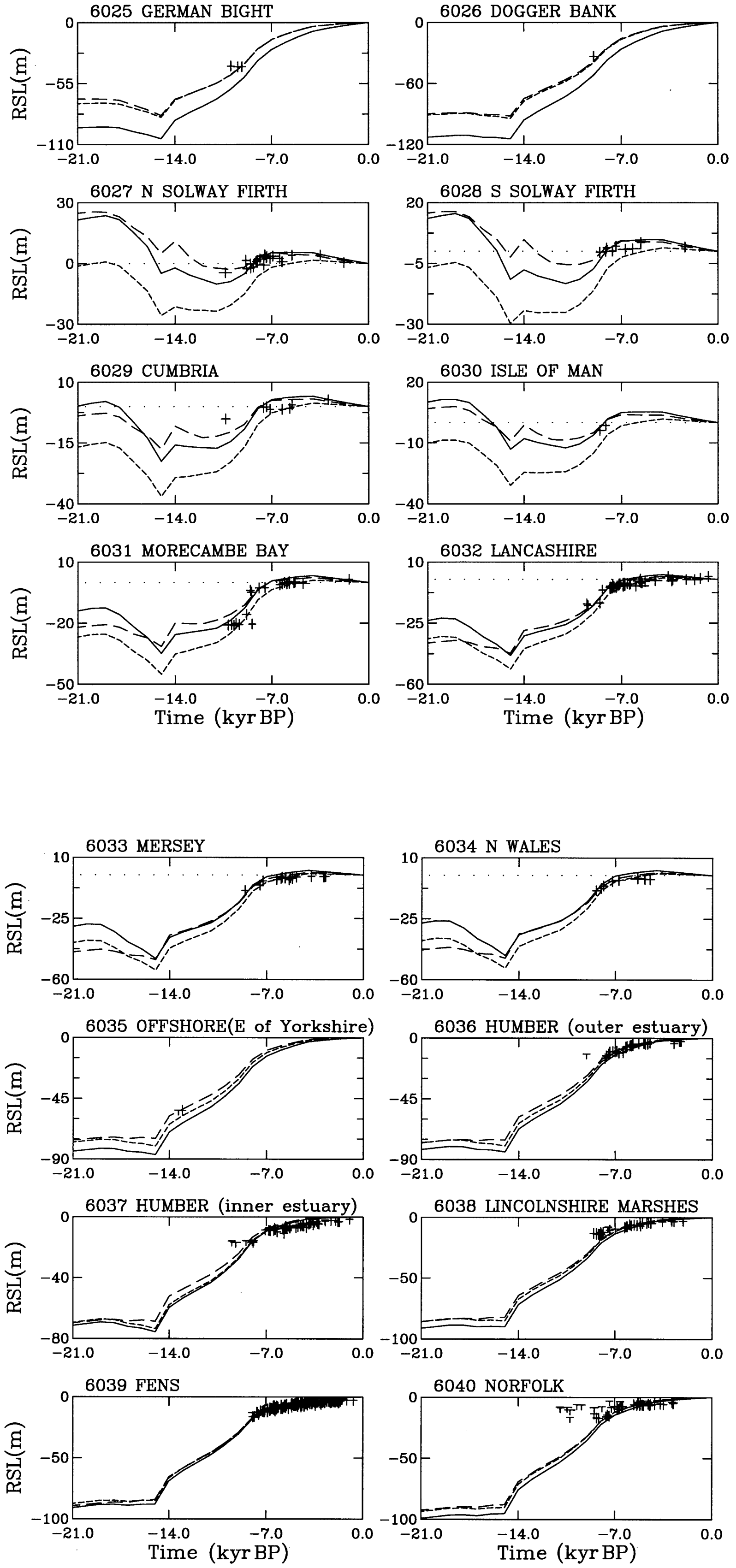


Figure 10. (*Continued.*)



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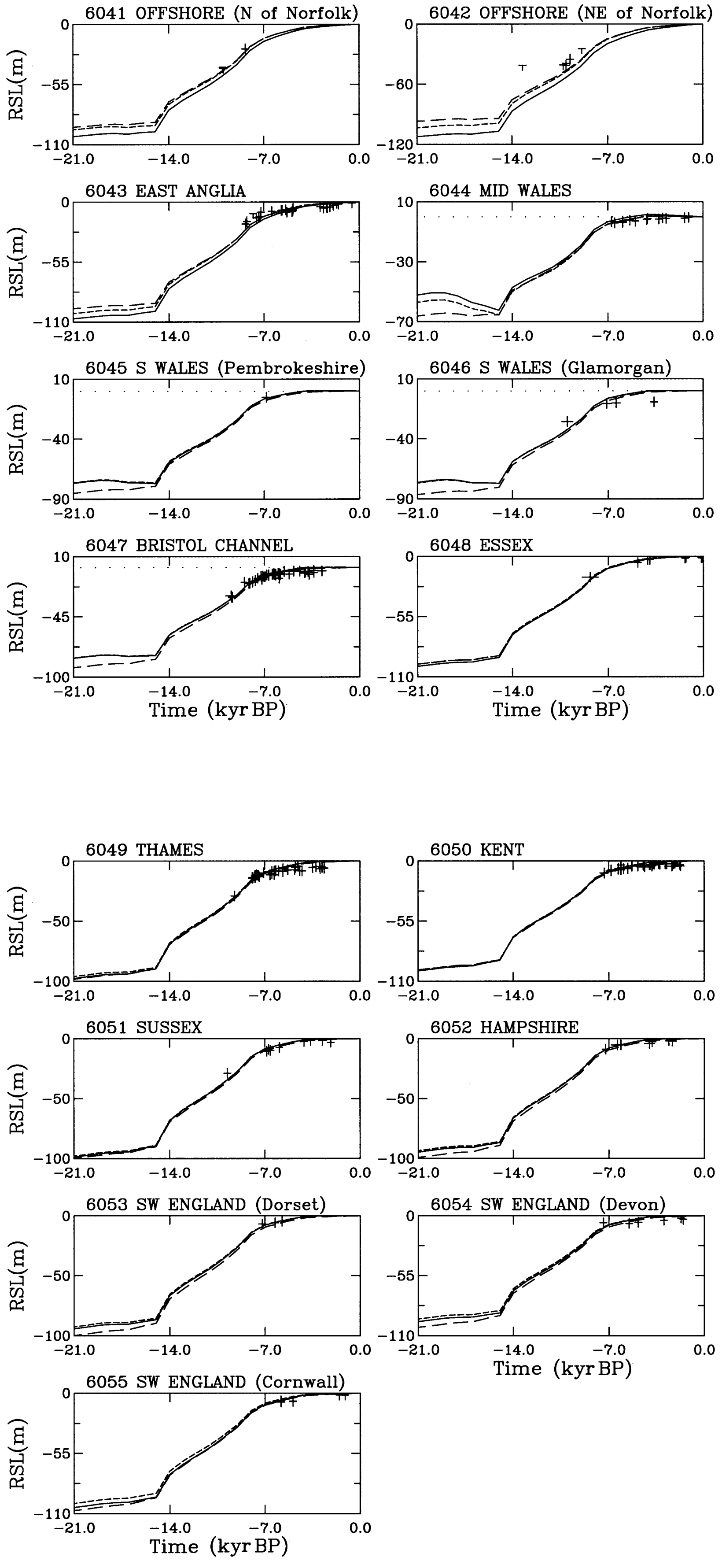

Figure 10. (*Continued.*)

which ice thickness and lithosphere thickness trade off in the RSL signature at Scottish sites is provided by the additional prediction shown as the dotted lines on the individual plates of Fig. 10. The predictions of this thin ice–thick lithosphere model clearly misfit the data at all northern sites to an entirely unacceptable degree.

Of the slight misfits of the theoretical predictions to the observations at sites that were near the southern boundary of the greatest concentration of land ice, those at the Morecambe Bay, Mersey and N. Wales locations are most pronounced. At these locations the theory predicts the occurrence of raised beaches that should extend to approximately 3 m above present sea level and have an age of 4 kyr; these are not observed. Since the model ice thickness for the Irish Sea and surrounding upland areas (NW England and Wales) is not constrained by observations comparable to those for NW Scotland by Ballantyne *et al.* (1998), the ice model (Fig. 8b) could simply be construed to have too much ice load for these areas. Explicit analyses of the sensitivity of this feature to the thickness of the local ice load (not shown) have, however, demonstrated that this potential problem cannot be resolved in this way.

Beyond this transition region, the fits of all of the models to the observations at locations in England are excellent. The abundant data set from the Fenlands (Fens), previously discussed with respect to Fig. 5, is extremely well reconciled, as are the data from Humber (inner and outer estuary locations), Lincolnshire Marshes, Norfolk and East Anglia. At the Mid-Wales site we observe the same characteristic misfit as occurs at the N. Wales location, the prediction here once more being characterized by a raised beach at 4 kyr which is ruled out by the observations. All of the time-series from locations still further south, such as those from the Bristol Channel, the Thames and Kent locations, from Sussex and Hampshire, Dorset, Devon and Cornwall, are nicely fit by all of the models, reinforcing once more the high quality of the basic ICE-4G (VM2) model which has been slightly perturbed in both its shallowest visco-elastic structure and in the maximum thickness of Scottish ice so as to enable the model to better satisfy the Ballantyne *et al.* (1998) constraint on the maximum thickness of the Scottish ice-sheet.

In the northernmost regions of Scotland, at Dornoch Firth, Moray Firth, the Hebrides and even the Tay and Forth Valley sites and at some sites further south, including NE England, Ayr and Tees, there is evidence that the degree of mid-Holocene non-monotonicity predicted using the VM2-L90 viscoelastic structure together with the thin-ice model is overly pronounced. We will return to a discussion of this issue in Section 4.3 below.

4.2 The influence of the Laurentide teleconnection on Scottish RSL histories

Clearly evident by inspection of all of the above RSL model predictions for Scottish sites that were covered by substantial thicknesses of glacial ice are two pronounced characteristics. First, and the one on which we will focus in this subsection, is that at all such sites in the time between 15 000 yr BP and 14 000 yr BP there is a predicted abrupt inflection or actual rise in RSL. This feature is associated with the rapid rise of global sea level that occurred during this period, as recorded in the sequence of corals at the island of Barbados (Fairbanks 1989; Bard *et al.* 1990), to which data set the ICE-4G (VM2) model was tuned in order to ensure that the eustatic sea-level history embodied within it was able to satisfy the global constraint provided by this unique record which extends back in time to the LGM itself. The rapid RSL rise feature that this record contains has been referred to as meltwater pulse 1-a, MWP1a, by Fairbanks (1989). Fig. 11 compares the sea-level history predicted by the ICE-4G (MV2) model to the Barbados observations, and demonstrates that the model properly includes this feature. The only site in Great Britain from which data are currently available that extend sufficiently far back in time to potentially record the impact of MWP1a is the Arisaig site in NW Scotland (Shennan *et al.* 1995); the data from this site are in fact of exceptionally high quality as they are based on the isolation basin method discussed in Section 2 of this paper and illustrated in Fig. 3. The data from Applecross and Kentra also provide some additional information for this critical range of time. Inspection of Fig. 9(b) demonstrates that the source of MWP1a in the ICE-4G reconstruction is associated primarily with an episode of rapid collapse of the North American ice complex.

Attention was first drawn to the fact that the ICE-4G (VM2) model predicts a rather pronounced signal at most Scottish sites due to MWP1a (see Fig. 10) in Peltier (1998a). At many of the northernmost sites, from which data and theoretical predictions are shown in Fig. 10, the predicted signal consists of an actual reversal of the general fall of relative sea level that occurs during this range of time due to the postglacial rebound of the crust. Shennan (1999) has recently considered the extent to which the data from Arisaig might be invoked either to deny or to confirm the possibility of such a reversal of sea-level tendency, and therefore the precise timing and magnitude of the meltwater pulse. If this prediction of a reversal of sea-level tendency due to the impact of MWP1a could be confirmed it would constitute the first truly independent test of its validity at any near-field location on the Earth's surface. This is especially important because the data are derived from land-based observations whose precision is not influenced by a range of factors which have an effect on the various coral-derived long records from the far-field sites of Barbados (Fairbanks 1989), the Huon Penninsula (e.g. Chappell *et al.* 1998) and Tahiti (Bard *et al.* 1996), and from the sediment-derived records from the Sunda Shelf (Hannebuth *et al.* 2000), the Argentinian shelf (Guilderson *et al.* 2000) and the northern Australian shelf (Yokoyama *et al.*

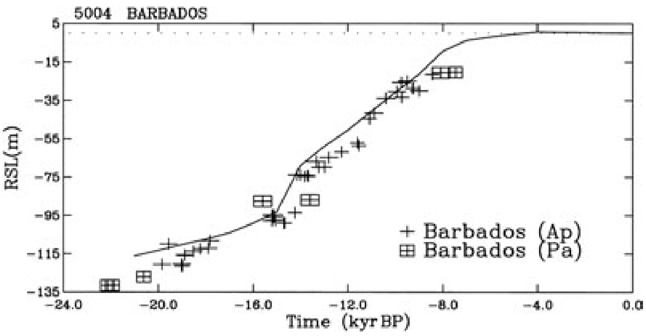

Figure 11. Comparison of the LGM to present RSL history observed at the Island of Barbados (Fairbanks 1989) with that predicted by the ICE-4G (VM2) model. The primary control points on the observed curve are determined by measurements or the coral species Acropora Palmata (Ap) which generally lives within 5 m of the level of the sea. Observations of sea level based upon the species Porites (Pa), which may live at much greater depth as shown in this Figure for several examples, cannot be employed to constrain sea level.

2000). These factors include episodic tectonic land movements, breaks in sediment accumulation or coral growth, lack of correlation between boreholes, large indicative ranges of the sample used for dating the SLIs, and sediment compaction. Both the Sunda shelf data and the Argentinian shelf data have been explicitly shown to require the presence of the same meltwater pulse 1a as was first revealed in the data from Barbados.

The analyses of Shennan (1999) were based on a comparison of the Arisaig data with the output of the models of Peltier (1998a) and various models of eustatic sea level, from Barbados (Fairbanks 1989; Fairbanks 1989), Tahiti (Bard *et al.* 1996) and those compiled and analysed by Fleming *et al.* (1998). In the predictions of RSL histories at Scottish sites by Shennan *et al.* (2000a; e.g. see Fig. 25 of that paper), the absence of any such reversed sea-level tendency feature associated with MWP1a results from using the 'nominal model' of eustatic sea-level history preferred by Fleming *et al.* (1998). Since this eustatic model contains no meltwater pulse feature there clearly can be no influence of any such feature transmitted into the local RSL history (Shennan 1999). Because the presence of MP1a is no longer debatable, based upon the recent publications of Hannebuth *et al.* (2000) and Guilderson *et al.* (2000), the model of eustatic sea-level history of Fleming *et al.* (1998) is no longer tenable. Shennan (1999) concluded that the current data from Arisaig cannot be construed to provide any support for the presence of an RSL reversal associated with MWP1a such as is predicted by the original ICE-4G (VM2) model in Peltier (1998a). The detailed stratigraphies in the isolation basins in this region strongly deny that any such event could have occurred, although there remains a small gap in the record from Arisaig that could accommodate a very small reversal. Shennan (1999) concluded that, even allowing for the small gap, MWP1a with a magnitude described by Fairbanks (1989) is not supported by the Arisaig data. However, the analysis in Shennan (1999) was dependent on the original ICE-4G deglaciation model for Britain (together with the original VM2 radial viscoelastic structure as discussed in the analyses of Peltier 1998a), a deglaciation model that does not satisfy the Ballantyne *et al.* (1998) constraints on Scottish ice thickness. It is therefore very notable that our thin lithosphere–thin ice model predicts no such reversal of sea-level tendency at Arisaig (Fig. 10), even though the ICE-4G global model of deglaciation contains a strong MWP1a pulse (Fig. 9). The only other locations with data for this period are Kentra and Applecross. Although limited to one or two dated SLIs, the isolation basins (Shennan *et al.* 2000a) from which these data come show no reversal of sea level at this time and therefore also support the RSL predictions from our thin lithosphere–thin ice model (Fig. 10).

These records (at Arisaig, Kentra and Applecross) have also played an extremely important role in the recent paper of Shennan *et al.* (2000a), in which it is strongly argued that the preferred shallow viscoelastic structure for Scotland should be one with a lithospheric thickness of $L=65$ km in order to best accommodate the Ballantyne *et al.* (1998) constraint on the maximum thickness of the Scottish ice-sheet as well as to fit the unique time-series from Arisaig. Their maximum ice-sheet thickness is very similar to the model we have optimized to work with our shallow viscoelastic structure with $L=90$ km, although the ice models have different spatial resolutions. The issue therefore clearly arises of why these two independent analyses lead to such discrepant conclusions. In our opinion a significant factor in explaining this discrepancy may lie in the fact that, in deciding upon the model that best fits the data, Shennan *et al.* (2000a) use the ^{14}C timescale throughout their analysis for RSL observations and ice model but not for the RSL predictions. However, we are not able to run model comparisons on the two timescales to confirm this explanation. Inspection of the ^{14}C timescale calibration shown in Fig. 3, however, demonstrates that the oldest data points at Arisaig have an age on the ^{14}C timescale that is younger by ~ 2.7 kyr than on the calendar year timescale. In order for our model to fit the data on this timescale we would have to thin the lithosphere further if all other properties of the model were held fixed. As we will see in what follows there are further, and most probably equally important, issues concerning the preference in the Lambeck *et al.* (1998) and Shennan *et al.* (2000a) analyses for the model with a lithospheric thickness of 65 km.

4.3 The influence of the Antarctic teleconnection on Scottish RSL history

The second of the observed characteristics of the RSL predictions at Scottish sites is much more pronounced even than that due to MWP1a, and is therefore at least equally interesting. This is that, at almost all such locations, the late Holocene is predicted to be characterized by falling sea levels associated with ongoing postglacial rebound of the once ice-covered crust, whereas during a restricted range of time within the early Holocene, from about 11 kyr BP to about 7 kyr BP, sea levels are observed to be rising, leading to highly non-monotonic records of RSL change at all Scottish locations. The explanation of this non-monotonicity of RSL history, as pointed out for example in Peltier (1998a), is well known to be due to the fact that the deglaciation of distant regions on the Earth's surface was proceeding sufficiently rapidly in this time interval that water was being added to the global ocean at a rate such that the local rebound of the crust in Scotland was unable to overcome the inundation of the land caused by the increasing eustatic rise. In the ICE-4G model of the global history of deglaciation, the source of the meltwater that allows the global theory to predict this feature of the observations in Scotland is primarily Antarctica. Inspection of Fig. 9(b) will demonstrate that it is in precisely this range of time that the partial deglaciation of West Antarctica and the loss of peripheral ice from East Antarctica is assumed to occur.

That it is in fact this single source of meltwater in the ICE-4G reconstruction that is primarily responsible for inducing the observed strong non-monotonicity of RSL history at the previously ice-covered sites in Scotland can be demonstrated explicitly by simply removing the Antarctic component of the melting history from the model and recomputing the solution to the integral sea-level equation. Fig. 12 shows comparisons at a series of sites in Scotland based on the use of the ICE-4G (VM2) model and for a perturbed model in which the Antarctic component of the melting history has simply been deleted. Also shown is the result for a model deglaciation history that has been slightly modified as shown by the dashed line on the eustatic curve for this region in Fig. 9(b). Clearly the complete removal of the delayed ice-sheet melting event for Antarctica from the ICE-4G model entirely eliminates the non-monotonicity from the predicted Scottish records. Now it might be imagined, and to some degree this is certainly the case, that the Scottish observations are not able to uniquely implicate an Antarctic

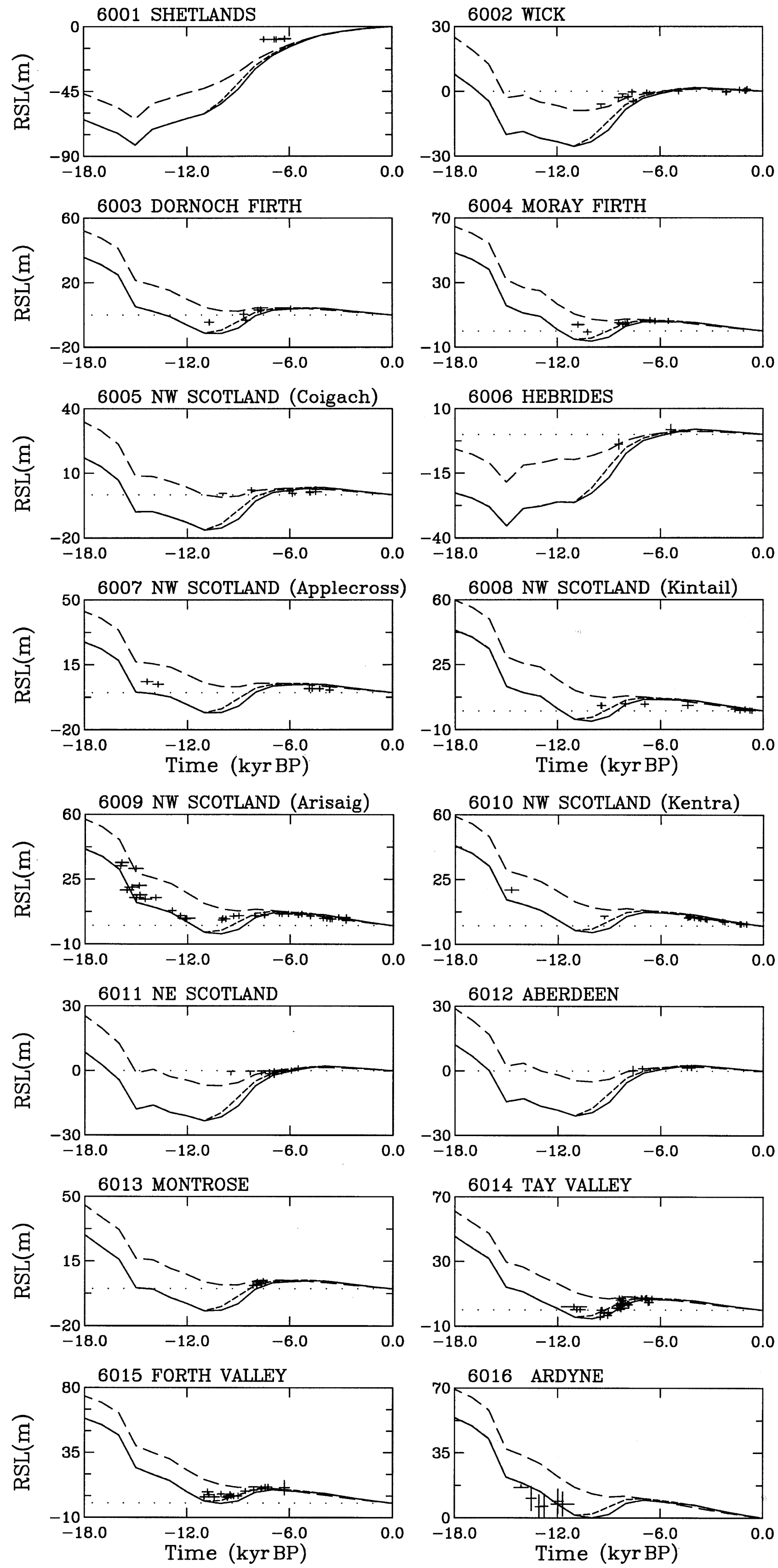

Figure 12. Comparison of observed RSL histories at selected British Isles locations with predictions based upon the thin ice–thin lithosphere model (solid lines) and with a model that is identical in every respect except that the contribution to the global deglaciation event from Antarctica is entirely removed (long-dashed lines). Also shown are predictions for the thin ice–thin lithosphere model for which the deglaciation event on Antarctica occurs somewhat earlier, as shown by the dashed form of the Antarctic contribution to eustatic sea-level history shown in Fig. 9.

source for the late intense melting event that is absolutely required to explain the observations from Scotland. However, it should be clear that these data do uniquely constrain the period of time within which additional water must be added to the global ocean so as to induce the several metres of sea-level rise that are observed so clearly to occur in this northern region of the British Isles because of the relative rates of local rebound of the crust and eustatic rise.

Because the local rate of crustal rebound, which is diminishing as a function of time, causes the amount of RSL rise that is actually observed to be less than the eustatic rise, it will be clear that the RSL rise during the early Holocene in Scotland will be a lower bound on the actual eustatic rise that must have been due to the distant melting event that generated the non-monotonicity. In the version of the ICE-4G model of global LGM to present deglaciation being employed herein, the eustatic sea-level rise induced by the partial deglaciation of Antarctica (see Fig. 9b) is approximately 17 m. If we were to imagine that a greater fraction of this eustatic rise were to have been derived from a North American source at this same time, which in ICE-4G is already contributing something to this event (see Fig. 9b), then we would probably be obliged to significantly modify the deep radial viscoelastic structure of VM2 in order to attempt to compensate for the increased postglacial rebound in the Hudson Bay region that would be produced. Since the relaxation times associated with the GIA process in this region are extremely well fit by the VM2 model (see Peltier 1998a for detailed discussion) there is good reason to doubt that a radical alternative of this kind to the ICE-4G deglaciation chronology is at all plausible. This being said, it should be clear that less dramatic modifications to the ICE-4G chronology are nevertheless allowed by the observations and will be required to improve the quality of the model. For example, many of the Scottish sites and those in NE England (Fig. 10) have the observations at *ca.* 8–7 kyr BP lying a little above the predicted RSL for the thin lithosphere–thin ice model, as well as a more modest amplitude of the period of mid-Holocene sea-level rise than that predicted. This suggests a slightly earlier melting of some of the Antarctic ice and/or some reduction in the amount of melting that was delivered by the West Antarctic Ice-Sheet (WAIS). As we will discuss in more detail elsewhere, all of these remaining misfits at Scottish and NE England sites may be entirely eliminated by appropriately modifying the history of Antarctic melting. Furthermore, recent efforts in the context of the QUEEN project (Svendsen *et al.* 1999) on the reconstruction of northwest European ice cover have reintroduced the idea that at the LGM there was indeed a connection between the Scottish and Fennoscandian ice-sheets, and have questioned the validity of the ICE-4G reconstruction insofar as the LGM margins along the southern flank of the Kara Sea are concerned. These are entirely legitimate issues that will be decided primarily on the basis of surface geomorphological observations rather than on the basis of the global viscoelastic theory of postglacial RSL change itself.

4.4 The influence on British Isles RSL histories of possible late Holocene instability of the Great Polar Ice-Sheets on Greenland and Antarctica

Because of the possible importance of modern ice-sheet melting to the ongoing global rate of RSL rise that is recorded on modern tide gauges (see Peltier 2000 for a detailed recent review), the issue of whether or not the existing great polar ice-sheets continued to lose mass subsequent to the mid-Holocene partial collapse of the Antarctic cryosphere is clearly important. As will be clear on the basis of the previously discussed Fig. 9(b), a fundamental assumption embodied in the (slightly modified) ICE-4G model of the last glacial–interglacial transition is that no further loss of mass occurred subsequent to approximately 4000 calendar years BP. In considering the validity of this assumption it is useful to return to the previously discussed comparisons of the RSL observations for the British Isles with the theoretical predictions for the sequence of models that differ only in their shallow viscoelastic structure (Fig. 10). More specifically, it is useful to focus on the sequence of sites close to the southern margin of the LGM ice-sheet (Lancashire, Morecambe Bay, Mersey, N. Wales and mid-Wales) where the theoretical models all predict the appearance of raised beaches through mid-Holocene time which are not observed at any but the Lancashire site. If there were to have been a continuing eustatic addition of water to the global ocean from about 4 kyr onwards, it should be clear by inspection of the misfits of the theoretical predictions to the data sets at these locations that they might be as easily eliminated by the action of such late Holocene melting from either Greenland or Antarctica as by modifying the local elements of the ice model as previously suggested. Since Fleming *et al.* (1998) have proposed that such late Holocene melting could well be occurring, it is important that we examine this possibility directly by appropriately further perturbing the ICE-4G model.

To this end we show in Fig. 13 a series of model predictions of the UK RSL time-series for the critical and proximate sites using our strongly preferred model with $L=90$ km, and for two variants of this model in which it has been assumed that additional melting from West Antarctica has been ongoing since 4 kyr BP at a sufficient rate to induce a continuing eustatic rise of sea level of either 0.25 mm yr^{-1} or 0.50 mm yr^{-1}, the latter rate having been suggested as an allowed upper bound on the late Holocene rate of eustatic rise in the recent paper by Fleming *et al.* (1998). These authors also identify an alternative eustatic curve that has 3.5 m of equivalent ice melt since 7 kyr BP, with most melting completed by 2.1 kyr BP. Since the predicted mid-Holocene highstands of sea level at Lancashire and other nearby sites are only about 2 m, it should be clear that, if an 'Antarctic-melting-tail' were of strength 0.5 mm yr^{-1}, continuing at this constant rate from 4 kyr BP onwards, the predicted highstands at these locations would be eliminated if the local rate of increase of bathymetry thereby induced were approximately equal to the globally averaged (eustatic) rate delivered by this modification to the base model. Inspection of Fig. 13 demonstrates that, although the mid-Holocene highstands are reduced in amplitude by incorporation of the late Holocene melting tail, perhaps sufficiently to render such a feature impossible to detect, the highstands still remain as features of the theoretical predictions. It will also be evident, however, that the influence of such incorporation of continuing late polar ice-sheet melting modifies the predicted RSL histories at other locations, in both Scotland and NE England, in a way that degrades the quality of the fits to the data of the model with $L=90$ km. It particular, the observed mid-Holocene emergence in NE England and SE Scotland is sufficiently reduced so as to render this modification to the eustatic curve unacceptable. The problem at these sites could most probably be rectified by a further thinning of the lithosphere. Since the analyses of Lambeck

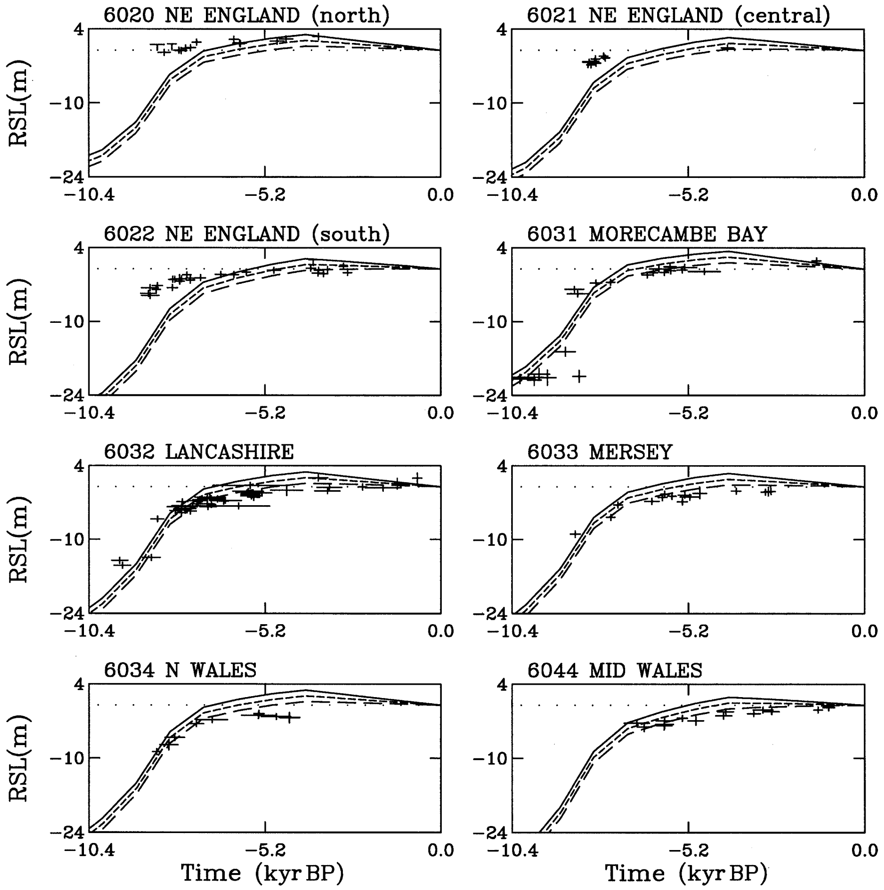

Figure 13. Comparisons of observed and predicted RSL histories at selected UK sites for a sequence of models that differ only in the melting histories assumed to have occurred over the past 4000 years. The predictions for the thin ice–thin lithosphere model are shown as the solid lines. In this model the melting of continental ice-sheets is assumed to have ceased entirely by 4000 years ago (see Fig. 9b). The short-dashed and long-dashed lines denote the RSL histories delivered by models that are otherwise identical to thin ice–thin lithosphere but which include, respectively, late Holocene 'melting tails' of respective strength 0.25 and 0.50 mm yr^{-1} which are assumed to derive from the melting of Antarctic ice.

et al. appear to have included late Holocene melting as in the Fleming *et al.* eustatic reconstruction, this may be a further reason why these authors have been led to suggest a model with a thinner lithosphere.

The best independent constraints available on the strength of the melting of polar ice that could have continued subsequent to mid-Holocene times consist of RSL histories from regions that are located as far as possible from the enhanced concentrations of land ice that were in place at the LGM. Furthermore, the data from oceanic island locations are especially useful in this regard as they are free of the influence of the hydroisostatic adjustment that occurs across continental coastlines in the far field of the ice-sheets due to tilting of the surface caused by the offshore weight of the water load that is emplaced upon the ocean basins by the deglaciation of the continents. At all such locations, a mid-Holocene 'highstand' of sea level is in fact ubiquitous. Fig. 15 shows an extensive sequence of comparisons for a large number of islands located at the sites in the equatorial Pacific Ocean denoted by the numbered locations in Fig. 14, at all of which direct evidence exists of the presence of the mid-Holocene highstand, at an elevation near 2 m at approximately 2–5 kyr BP. The highstand observations being employed for the purpose of these comparisons are those recently compiled and quality controlled in the paper by Grossman *et al.* (1998). Evident upon comparison of the set of three model predictions with the observations at each site is the fact that our preferred model with $L=90$ km predicts the altitude of the highstand extremely well at every location. In comparison, the model that includes an 'Antarctic melting tail' of strength 0.25 mm yr^{-1} underpredicts the amplitude of the highstand by approximately 1 m, whereas the model with a tail strength of 0.5 mm yr^{-1} eliminates the highstand completely. We simply note here (results not shown) that the same result is obtained when the late Holocene melting is assumed to derive from Greenland.

The idea that a late Holocene melting tail of the eustatic sea-level history of the magnitudes described above might be invoked to explain the misfits at locations such as Lancashire, Mersey and N. and S. Wales in the British Isles data set is therefore ruled out by the requirement that the model be simultaneously compatible with the global data set. We argue that an Antarctic melting tail that continues to the present day, with an upper limit in strength of 0.5 mm yr^{-1}, similar to that suggested as possible in the nominal eustatic model described by Fleming *et al.* (1998), is not supported by our analysis. In order to achieve their best fits, with respect to both altitude and duration, to the Holocene highstands at sites in NW Scotland, Shennan *et al.* (2000a) required a modification to the nominal eustatic model suggested by Fleming *et al.* (1998). Specifically, they applied an *ad hoc* 'correction' of 3 m of melting since

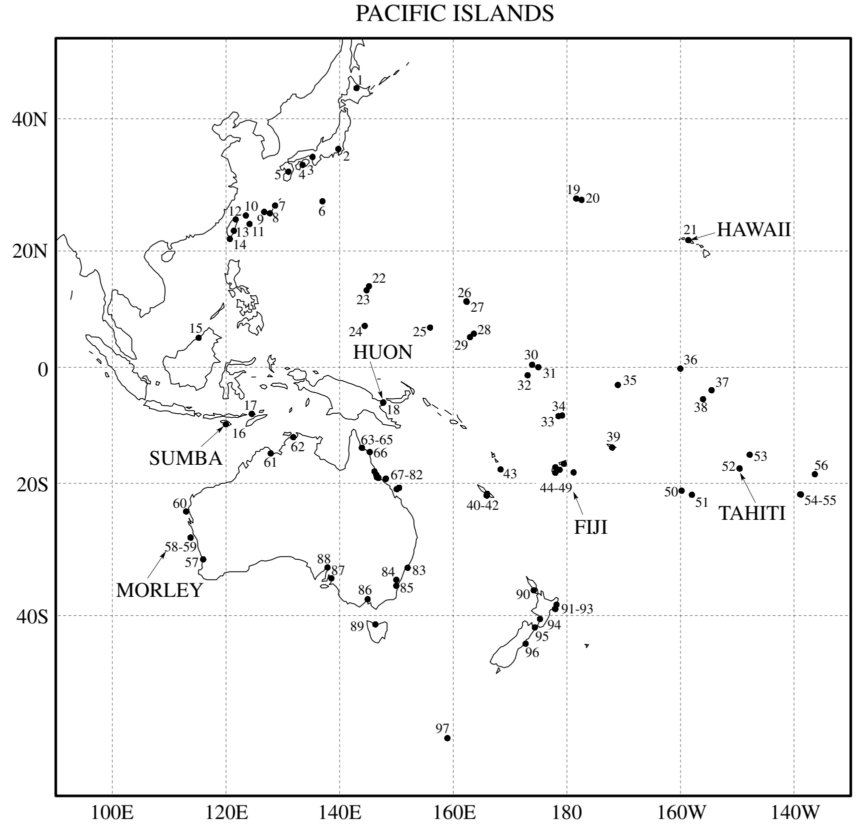

Figure 14. Location map for the equatorial Pacific islands from which shoreline data are available which record the mid-Holocene highstand of sea level which may be invoked to rule out the possibility of any significant melting of the great polar ice-sheets during the late Holocene period.

6 ^{14}C kyr BP, with most ($\sim$2.5 m) completed by 2 ^{14}C kyr BP, and all by 1 ^{14}C kyr BP. Although this *ad hoc* correction, which was not attributed to melting from any specific location, still gave RSL predictions lying above the observations, Shennan *et al.* (2000a) simply defined this as the upper limit of melting that would not remove the mid-Holocene highstands at far-field locations, essentially the same constraint that we have discussed herein (see also Peltier 2000). In summary, the eustatic model (equivalent ice-melt history) for the mid- and late Holocene employed by Shennan *et al.* (2000a) is very similar to that used here (Fig. 9b), both with $\sim$3 m of eustatic rise since 6 kyr BP and declining approximately exponentially afterwards, but ending at 1 ^{14}C kyr BP in the case of Shennan *et al.* (2000a) and by 4 kyr BP in the present case.

4.5 A refined model designed to fit the British Isles data set and its global acceptability

Based upon the analyses discussed in the previous subsections of this paper, in order to best fit the observed RSL histories from the previously ice-covered region of Scotland and to satisfy the recent Ballantyne *et al.* (1998) constraint on maximum LGM ice-thickness over this region, a model with a value of lithospheric thickness L equal to approximately 90 km and with somewhat higher sublithospheric viscosity than exists in the transition region and upper mantle, such as exists in VM2, is rather strongly preferred. As illustrated in Fig. 7, this model also fits the relaxation spectrum for the Fennoscandian rebound of McConnell (1968) at the longest wavelengths, if lying somewhat towards the lower viscosity boundary of acceptability over the range of spherical harmonic degrees from $l=14$ to $l\approx40$ in which the relaxation times are well constrained according to the recent re-analysis of this spectrum by Wieczerkowski *et al.* 1999). At the shortest wavelengths, however, it deviates substantially from the McConnell (1968) spectrum, as allowed by the fact that the relaxation times at short wavelengths are essentially unconstrained. Since the viscosity in the lower part of the lower mantle of the preferred model has not been modified at all from VM2, and since this has been constrained in the construction of VM2 (Peltier & Jiang 1996; Peltier 1996a, 1998a) entirely by the observation of the non-tidal acceleration of planetary rotation (or equivalently the present-day time derivative of the degree-2 axial component of the spherical harmonic expansion of the planetary gravity field), it should be clear that the revised model will continue to satisfy this constraint on the viscosity of the deepest part of the lower mantle. Furthermore, since the viscosity of the upper part of the lower mantle is constrained by the observed relaxation times that characterize the postglacial rebound process in the Hudson Bay region (e.g. see Peltier 1998a), it will be clear, since the viscosity over this depth range has not been modified, that the refined model will continue to fit these relaxation times.

In fact, the only data whose satisfactory explanation could conceivably be at risk because of the thinner lithosphere are those data from the US east coast that were found to be reconciled so well by the ICE-4G (VM2) model with a 120.7 km thick lithosphere, and other data that characterize the rebound process of crust that was once loaded by ice-sheets of similarly small scale to that which covered Scotland and which will therefore also be sensitive to this feature of the model. Sensitivities

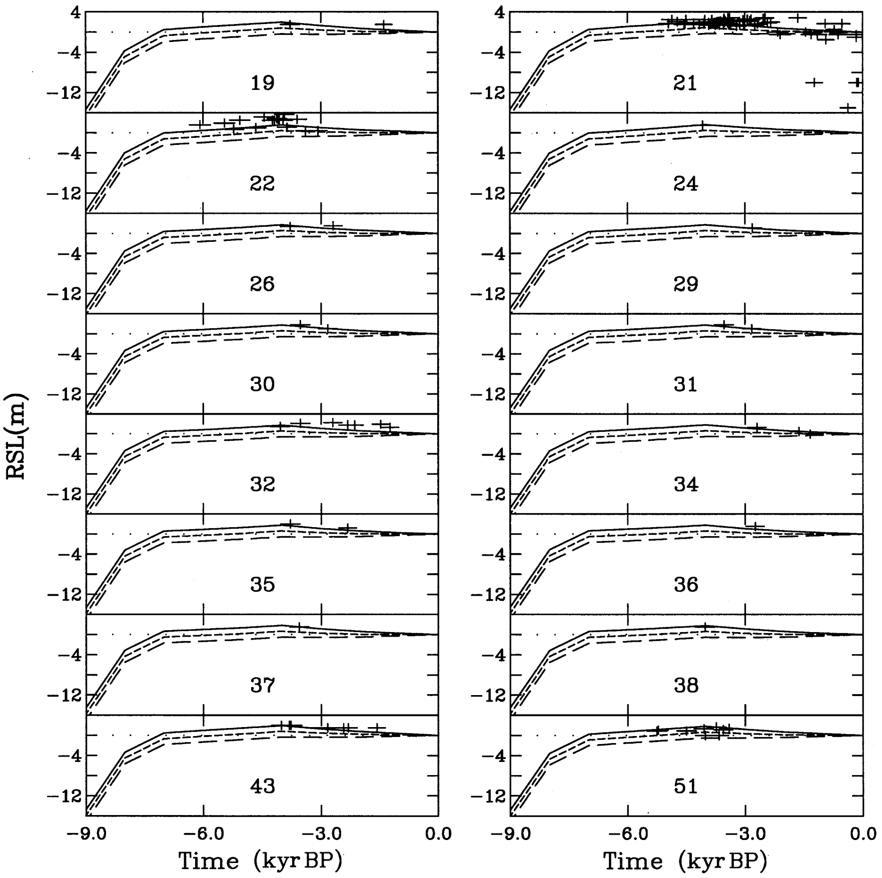

Figure 15. Comparisons of observed and predicted RSL histories at the equatorial Pacific island locations shown in Fig. 14. The solid curves denote the predictions made using the thin ice–thin lithosphere model in which all ice-sheet melting is assumed to have ceased by 4 kyr ago. The long-dashed and short-dashed curves denote the predictions made by the models which include 'melting tails' derivative of continuing Antarctic deglaciation of respective eustatic strength 0.25 and 0.5 mm yr^{-1}. It will be noted that, although the standard model reconciles the observations extremely well, neither of those that include a significant Antarctica-derived melting tail is able to fit the data.

of the latter kind, evident at sites in the Barents and Kara seas, the Queen Elizabeth archipelago of Arctic Canada (which was covered by the small-scale Innuitian ice-sheet) and perhaps also on the west coast of North America at sites that were under the Cordilleran complex, may be simply eliminated by a slight reduction of the local thickness of glacial ice and /or by altering the timing of deglaciation itself (where this has been imperfectly constrained). Because of the importance of RSL records from the US east coast, however, it is especially important that we demonstrate explicitly that these records are not violated by the thinning of the lithosphere that is required by the British Isles data. At these locations, adjustments of the loading history cannot be invoked to significantly ameliorate any misfits that might arise, as the thickness of the Laurentide ice-sheet over Hudson Bay is well constrained by the amplitude of emergence that is observed to have occurred there and because this amplitude is essentially independent of lithospheric thickness because of the extremely large horizontal scale of the ice-sheet.

In order to demonstrate the extent to which the model refined on the basis of the British Isles data set is acceptable to North American east coast observations, consider the results of the comparisons documented in Fig. 17 at the sites shown in Fig. 16(a). At the SE Hudson Bay location, for which the complete set of ^{14}C data was recently compiled in Peltier (1998a), it will be clear that each of the two models for which predictions are shown, which differ only in terms of lithospheric thickness and UK ice load, makes an essentially identical prediction of RSL history and each fits the observations essentially perfectly. The maximum thickness of the Laurentide ice-sheet is therefore extremely well constrained by the maximum emergence that characterizes such rebound curves because the viscosity of the upper part of the lower mantle is essentially independently determined by their relaxation times. The remaining sites for which comparisons of theory and observations are shown in this Figure are for locations that extend from the northern part of the east coast of the US to the south, beginning in the north at Boston. It will be clear that at all such locations the differences between the predictions of the two models and the observations are small but do vary systematically along the coast such that significantly lower (slightly higher) rates of present-day sea-level rise are predicted along the southern (northern) part of the coast for the model with $L=90$ km than for the original ICE-4G (VM2) model in which the lithospheric thickness is $L=120.7$ km. It is also evident, however, that the new model with $L=90$ km further and significantly improves the fit of the theoretical predictions to the RSL observations at the southernmost locations where its influence is most pronounced. We are therefore in a position to claim that the revised shallow viscoelastic structure which has $L=90$ km but VM2 sublithospheric viscosity is not significantly inferior to the original model, insofar as the data from the US east coast region are concerned.

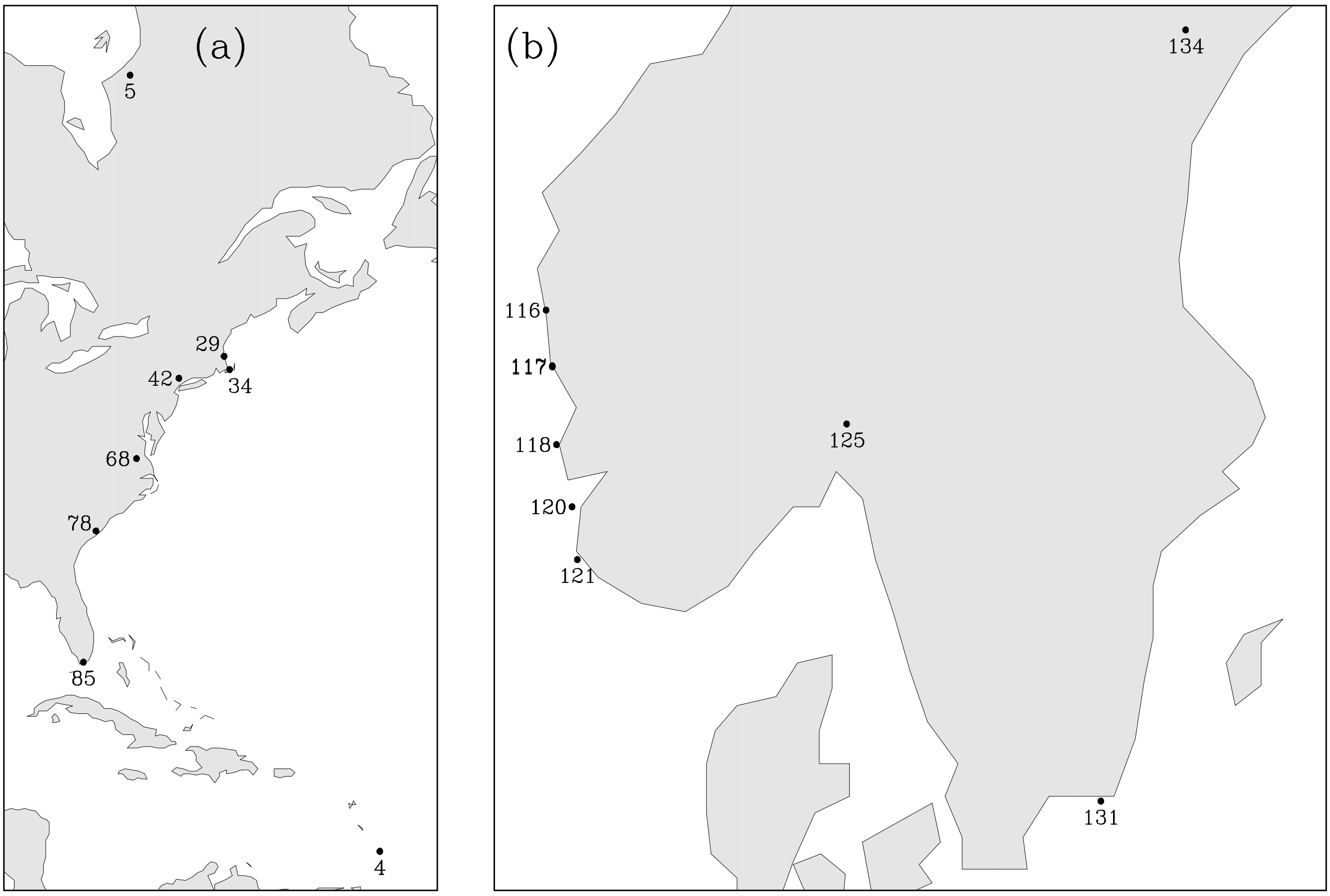

Figure 16. Location maps for sites at which comparisons are shown between theoretical predictions for the ICE-4G (VM2) and thin UK ice–thin lithosphere model with the observational data at sites from North America (a) and Fennoscandia (b).

A similar set of comparisons for the sequence of Fennoscandian sites with locations shown in Fig. 16(b) is shown in Fig. 18. For the purpose of these comparisons, no attempt has been made to modify the history of Fennoscandian deglaciation so as to correct any of the misfits that arise due to the use of the model with thinner lithosphere. All calculations therefore have essentially the same ice-thickness distribution and time dependence as in ICE-4G, although modest changes have been introduced in peripheral regions to the west and southwest of the core of the ice-complex (see Fig. 8). Included in this Figure are comparisons at Angermanland, Sweden, the 'classical' location from a site that was under the centre of the Fennoscandian ice-sheet and from which an extremely high-quality data set is available. Very careful inspection of the predictions at this site will show that the thinner the lithosphere in the model, the greater the amount of uplift predicted, although the impact upon the theoretical prediction is not nearly so severe in a relative sense as for the data from Scotland. The misfit induced by the model with thinner lithosphere could be simply corrected by thinning the Fennoscandian ice-sheet by about 10 per cent, very much less proportionally than the factor of 2 by which the Scottish ice-sheet had to be thinned to accommodate the shift from the model with a lithospheric thickness of 120.7 km to that with a lithospheric thickness of 90 km. At the other Fennoscandian sites for which comparisons are shown in Fig. 18, most of which consist of edge sites at which the data show evidence of the same non-monotonicity of the sea-level record as previously discussed based on the data from Scottish sites, the impact is more pronounced but is of a nature such as will be easily corrected by a slight thinning of the proximate ice load.

5 CONCLUSIONS

The analysis of postglacial relative sea-level histories from the British Isles discussed in the preceding sections of this paper have led us to the primary conclusion that the recently published constraint on the maximum thickness of the LGM ice-sheet over Scotland (Ballantyne *et al.* 1998) requires that the lithospheric thickness assumed in the ICE-4G (VM2) model be reduced from approximately 120 km to approximately 90 km. This allows the maximum LGM thickness of the Scottish ice-sheet to be reduced from a value near 2200 m to a value near 1200 m, the same ice-sheet thickness as has recently been found to be acceptable on the basis of the same criterion by Shennan *et al.* (2000a). However, in the Shennan *et al.* (2000a) analysis the Ballantyne *et al.* constraint is met only by assuming a lithospheric thickness of 65 km. The primary reasons for the difference in these two inferences of the thickness of the lithosphere appear to be associated with three key differences in the methodology employed to compare the observations with the predictions of the theoretical model. The first of these differences is that Shennan *et al.* (2000a), as in earlier analyses using the same approach (Lambeck 1993a,b, 1995, 1998; Lambeck *et al.* 1996), perform calculations in which the ice-sheet deglaciation history and the RSL observations on the

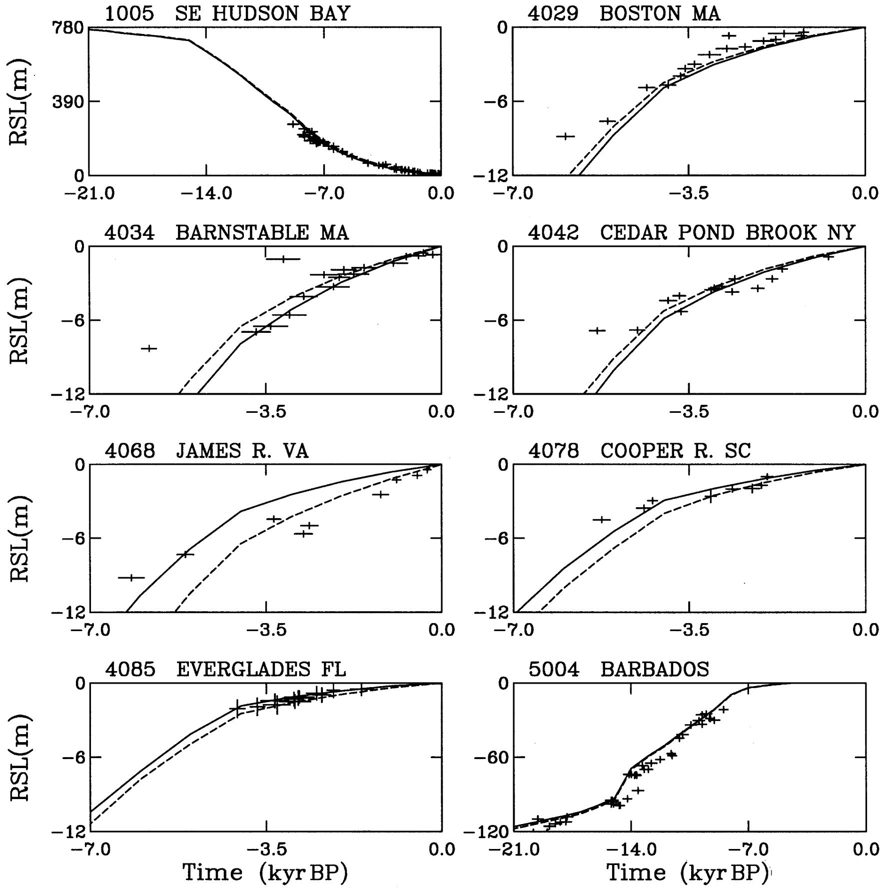

Figure 17. Comparisons of RSL observations and model predictions at six north American sites for the standard ICE-4G (VM2) model with $L = 120$ km (solid lines) and for the model with $L = 90$ km (short-dashed lines).

^{14}C timescale are compared with theoretical predictions on the 'Newtonian' calendar year timescale. This would be a source of error in the inference of Earth structural properties unless it were possible to precisely compensate for this inconsistency by assuming the depth-dependent viscosity structure inferred by fitting the model to the observations to be one measured in Pa ^{14} Cs, as suggested in Lambeck (1998). This is clearly not possible for rebound data from a region such as Scotland where the data are as sensitive to lithospehric thickness as they are to mantle viscosity and where the character of the RSL curves is complex. Second, there are also differences in the earth and ice models in terms of the time-steps employed, the spatial resolution, and the radial zonation of the viscosity structure. The separate effects of the latter influences are difficult to isolate but the most important of them is probably that due to the use of a simple three-layer viscoelastic structure (lithosphere to 65 km, upper mantle of 4×10^{20} Pa ^{14}Cs to 670 km, lower mantle of 10^{22} Pa ^{14}Cs to the core–mantle boundary) by Shennan *et al.* (2000a) compared with the models we have employed (Fig. 7) in which the viscosity varies more smoothly with depth. Lambeck *et al.* (1996) did not examine smoothly varying models of the radial viscoelastic structure in their work [upon which the three-layer model used by Shennan *et al.* (2000a) is based]. Likewise, Lambeck *et al.* (1996) did not have available the large number of new SLIs that have recently been published as output from the LOIS project. The existence of properly dated SLIs for the 16–9 kyr BP range of time in particular, rather than estimated ages of shoreline features that had not been directly dated, could well have a significant effect on minimum variance calculations to assess the goodness of fit of different models.

It appears to be the case, however, that the primary explanation for the significantly different model of the shallow viscoelastic structure that we have inferred herein has to do with the very strong impact upon the details of the RSL predictions at Scottish sites produced by the influence of global teleconnections due to the details of the deglaciation process that is assumed to be occurring at distant locations. Specifically, models that assume that more meltwater is added to the oceans during the Holocene period than assumed in ICE-4G, such as appears to be the case in the model analysed by Lambeck and others, will inevitably lead to the requirement for reduced lithospheric thickness in order to deliver sufficient postglacial rebound to fit the observations. The importance of such teleconnections due to distant ice-sheet melting to the interpretation of the isostatic adjustment process experienced by the British Isles has never been considered in any previous work. The explicit analyses reported herein show that these influences are vitally important. We have specifically drawn attention herein to the fact that the eustatic function employed in the analyses by Lambeck and co-workers and discussed in Fleming *et al.* (1998) is untenable in detail as it fails to incorporate the MWP1a feature whose presence has recently been verified in the analysis of Sunda Shelf data in Hannebuth *et al.* (2000) and because this eustatic function has been assumed to incorporate a late Holocene melting 'tail' of a strength that appears to be

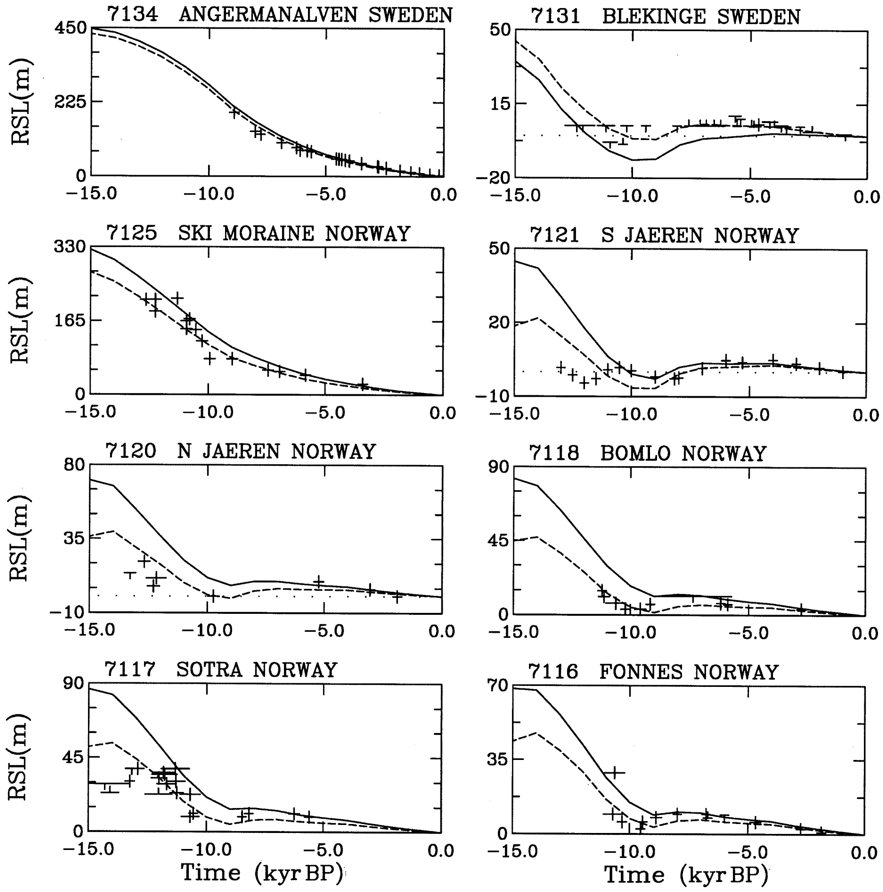

Figure 18. Comparisons of RSL observations and model predictions at six European sites for the standard ICE-4G (VM2) model with $L = 120$ km (solid line) and for the model with $L = 90$ km (short dashed lines).

ruled out by far-field observations of the mid-Holocene highstand. If this group were to publish the detailed space–time history of the deglaciation event employed in their models, which it has never done in the past, it would be possible to further explore the reasons for our differences of interpretation.

The trade-off between lithospheric thickness and the viscosity profile of the upper mantle, together with the sensitivity of RSL response to the details of the British ice-sheet load, has been noted in all of these previous studies and our analysis confirms this. What we have added herein is a detailed discussion of the sensitivity of the interpretation of these data to the assumptions made regarding the eustatic sea-level curve employed to describe the global declaciation process. Given the great increase in high-quality RSL data published since 1996, especially for the period prior to 10 kyr BP, against which we have tested the current model, and the use of the calendar year timescale throughout, we strongly favour a value of ~ 90 km for lithospheric thickness in this geographical region. This value is perfectly consistent with the entirely independent estimate based upon the re-analysis of ocean-floor bathymetry and gravity data recently reported by DeLaughter *et al.* (1999), and with the thermal boundary layer thickness to which this result refers. By revisiting the original analyses of such data reported by Parsons & Sclater (1977), and by employing the more extensive data set that has become available since that time, DeLaughter *et al.* (1999) have been led to a revised estimate of the asymptotic lithospheric thickness appropriate for the oldest ocean floor. Whereas the Parsons & Slater (1977) analysis gave an estimate of 125 ± 10 km for this thickness, the re-analysis of DeLaughter *et al.* (1999) has led to a downward revision to a value of 95 ± 10 km, a value with which our analysis of postglacial RSL data from the British Isles is in accord. This new value for lithospheric thickness will be further tested prior to accepting it as a feature of the ICE-5G (VM3N) model of the last deglaciation event of the current ice-age that is now in the final stages of development.

In demonstrating the extreme sensitivity of the inference of the shallow viscoelastic structure of the Earth, using glacial isostatic adjustment data for the British Isles, to the details of the global eustatic curve of the model (see Peltier 2000 and Shennan *et al.* 2000 for further discussion), we are in a position to reconsider conclusions reached previously on the basis of the analysis of data from the same region. Specifically, in Lambeck *et al.* (1996, page 353) it is asserted that 'Likewise, the imposition of a thick lithosphere on the viscosity structure, without conducting a full search of the model space, leads to a local minimum solution that does not correspond to the global minimum variance. Thus, the adoption of a relatively thick lithosphere, of say 100–120 km, as used in the recent models of Peltier and colleagues, leads to an over-estimation of the upper mantle viscosity and to models in which the contrast between upper and lower mantle viscosity is much less than found here (see, e.g. Fig. 2c)'. It is clear on the basis of the analyses presented herein, on the contrary, that the inference in Lambeck *et al.* (1996) of an extremely thin (65 km) lithosphere and a high upper mantle–lower mantle viscosity contrast was strongly influenced by their (questionable) assumptions concerning post-LGM eustatic sea-level history. Because their minimum variance solution was constructed without regard to this dependence or to the dependence upon the timescale on the basis of which the analysis of variance was performed, this led to a gross distortion of the values inferred for Earth structural parameters such as lithospheric thickness and sublithospheric viscosity.

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