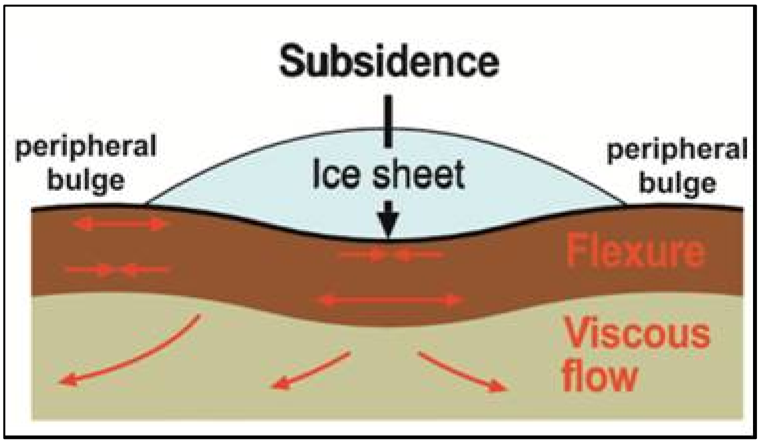
Potential Natural Changes and Implications for a UK GDF

Minerals and Waste Programme

Commissioned Report CR/12/127N



British Geological Survey
NATURAL ENVIRONMENT RESEARCH COUNCIL

British Geological Survey
Environmental Science Centre
Keyworth
Nottingham
United Kingdom
NG12 5GG

Tel +44 (0)115 936 3143
Email enquiries@bgs.ac.uk

REPORT APPROVAL FORM

Report title and authors

Potential Natural Changes and Implications for a UK GDF

Shaw, RP, Auton, C A, Baptie, B, Brocklehurst, S, Dutton, M, Evans, DJ, Field, LP, Gregory, SP, Henderson, E, Hughes, AJ, Milodowski, AE, Parkes, D, Rees, JG, Small, J, Smith, N, Tye, A, and West, JM

Client name and address	
NDA-RWMD Curie Avenue Harwell Didcot Oxfordshire OX11 0RA	BGS Report No. CR/12/127N BGS project code NEE4118 Classification None

	Name	Signature	Date
Prepared by	**RP Shaw**		**12th March 2013**
Checked by	**FM McEvoy**		**13th March 2013**
Approved by	**AJ Bloodworth**		**14th March 2013**

BRITISH GEOLOGICAL SURVEY

Minerals and Waste Programme

COMMISSIONED REPORT CR/12/127N

Potential Natural Changes and Implications for a UK GDF

Shaw, RP, Auton, C A, Baptie, B, Brocklehurst, S, Dutton, M, Evans, DJ, Field, LP, Gregory, SP, Henderson, E, Hughes, AJ, Milodowski, AE, Parkes, D, Rees, JG, Small, J, Smith, N, Tye, A, and West, JM

Author affiliations:

British Geological Survey (Shaw, RP, Auton, C A, Baptie, B, Evans, DJ, Field, LP, Gregory, SP, Hughes, AJ, Milodowski, AE, Parkes, D, Rees, JG, Tye, A, West, JM)

Natioanl Nuclear Laboratory (Dutton, M, Henderson, Small, J, Smith, N)

University of Manchester (Brocklehurst, S)

Keywords

Future Natural Change, GDF, seismicity, Glaciation, Weathering, Erosion, Uplift, Subsidence.

Front cover

See Figure 5 for details.

Bibliographical reference

Shaw, RP, Auton, C A, Baptie, B, Brocklehurst, S, Dutton, M, Evans, DJ, Field, LP, Gregory, SP, Henderson, E, Hughes, AJ, Milodowski, AE, Parkes, D,Rees, JG, Small, J, Smith, N, Tye, A, and West, JM 2012. Potential Natural Changes and implications for a UK GDF. *British Geological Survey Commissioned Report*, CR/12/127. 198pp.

Keyworth, Nottingham British Geological Survey 2012

BRITISH GEOLOGICAL SURVEY

The full range of our publications is available from BGS shops at Nottingham, Edinburgh, London and Cardiff (Welsh publications only) see contact details below or shop online at www.geologyshop.com

The London Information Office also maintains a reference collection of BGS publications, including maps, for consultation.

We publish an annual catalogue of our maps and other publications; this catalogue is available online or from any of the BGS shops.

The British Geological Survey carries out the geological survey of Great Britain and Northern Ireland (the latter as an agency service for the government of Northern Ireland), and of the surrounding continental shelf, as well as basic research projects. It also undertakes programmes of technical aid in geology in developing countries.

The British Geological Survey is a component body of the Natural Environment Research Council.

British Geological Survey offices

BGS Central Enquiries Desk

Tel 0115 936 3143 Fax 0115 936 3276

email enquiries@bgs.ac.uk

Environmental Science Centre, Keyworth, Nottingham NG12 5GG

Tel 0115 936 3241 Fax 0115 936 3488
email sales@bgs.ac.uk

Murchison House, West Mains Road, Edinburgh EH9 3LA

Tel 0131 667 1000 Fax 0131 668 2683
email scotsales@bgs.ac.uk

Natural History Museum, Cromwell Road, London SW7 5BD

Tel 020 7589 4090 Fax 020 7584 8270
Tel 020 7942 5344/45 email bgslondon@bgs.ac.uk

Columbus House, Greenmeadow Springs, Tongwynlais, Cardiff CF15 7NE

Tel 029 2052 1962 Fax 029 2052 1963

Maclean Building, Crowmarsh Gifford, Wallingford OX10 8BB

Tel 01491 838800 Fax 01491 692345

Geological Survey of Northern Ireland, Colby House, Stranmillis Court, Belfast BT9 5BF

Tel 028 9038 8462 Fax 028 9038 8461

www.bgs.ac.uk/gsni/

Parent Body

Natural Environment Research Council, Polaris House, North Star Avenue, Swindon SN2 1EU

Tel 01793 411500 Fax 01793 411501
www.nerc.ac.uk

Website www.bgs.ac.uk
Shop online at www.geologyshop.com

This report has been prepared by the British Geological Survey under contract to NDA. The report has been reviewed by NDA and Quintessa on behalf of the NDA, but the views expressed and conclusions drawn are those of the authors and do not necessarily represent those of NDA.

Conditions of Publication

Bibliography

If you would like to see other reports available from the NDA, a complete listing can be viewed at our website www.nda.gov.uk, or please write to the Communications Department at the address below.

Feedback

Readers are invited to provide feedback to the NDA on the contents, clarity and presentation of this report and on the means of improving the range of the NDA reports published. Feedback should be addressed to:

Dr Elizabeth Atherton
Head of Stakeholder Engagement and Communications
Nuclear Decommissioning Authority
Radioactive Waste Management Directorate
Building 587
Curie Avenue
Harwell Oxford
Didcot
Oxfordshire OX11 0RH

e-mail rwmdfeedback@nda.gov.uk

Executive Summary

A period of one million years following closure has been used by RWMD when considering the post-closure safety case for a geological disposal facility (GDF). It is during this period that evolution of the near-field and local geosphere as a result of GDF construction and operation will be at its most rapid and radioactivity of the emplaced waste will be at their highest levels. Significant effort has been spent internationally on identifying the many natural processes that may affect the evolution of the geosphere over this timescale and the contribution of those processes to GDF performance. The purpose of this report is to identify which processes are relevant to geosphere evolution in this time period around a generic GDF in the UK. Previous work has identified tectonic effects, climate change effects, uplift, subsidence, volcanism and diagenesis as key concerns. The potential impact of each of these processes on a generic UK GDF, constructed according to a multiple barrier concept and sited at a depth of between 200 and 1000 m in a suitable host rock, is outlined in the following sections: tectonic related uplift and subsidence; seismicity, tectonic history and volcanism; climate change and glaciation and weathering and erosion.

GDF EVOLUTION

In the period following closure, the waste packages will largely contain the waste for the order of the first thousand years, during which time short-lived radionuclides will largely have decayed and the overall radionuclide inventory will be significantly reduced. Over a period of the order of one hundred thousand years, some components of the engineered system may degrade. During this period, the evolution of the geosphere is particularly important because it provides isolation and containment in conjunction with the backfill and buffer. The geosphere then continues to fulfil an isolating role and acts to contain some of the small proportion of radioactivity that has migrated from a GDF.

GROUNDWATER FLOWS

Changes to groundwater flow around a GDF are likely to be influenced by tectonic processes, glacial processes, long-term climate change, erosion and river capture. Often multiple associated processes will combine to affect groundwater flow. For example, during glaciation, changes to groundwater flow will result from an interaction of changes in pressure caused by the mass of ice and/or glacial melting, associated erosional processes and changes to sea-level which, together, are likely to affect the flow of groundwater around a GDF. Permafrost to depths of several hundred metres is likely to occur during future cold climatic phases. Permafrost may affect both the engineered barrier system/wasteform and host rocks. As it forms, solutes will remain in solution leading to the development of brines. Frozen ground will act as a barrier to groundwater flow while remaining frozen, but may have enhanced permeability once thawed causing temporary or more permanent changes to groundwater flow paths. Depending on depth, temperature and the presence of gases, in particular methane and carbon dioxide, permafrost conditions may lead to the formation of gas hydrates which will melt rapidly when temperature increases or pressure decreases. This will prevent gas migration when present but may result in rapid movement when melting happens. This combination of eustatic sea-level changes, changes to the hydrological system and changes to the rate and form of erosion and deposition suggest that, over the million year period considered, climate change has the potential to be the greatest influence on processes driving radionuclide migration and associated biosphere impacts.

Evaporites in general, but salt in particular, are soluble in fresh water. If groundwater flow patterns are changed by any processes to the extent that they are in direct contact with fresh

groundwater then their dissolution will occur. This is likely to have a significant effect on a GDF hosted in such rocks.

TECTONIC RELATED UPLIFT AND SUBSIDENCE

There are two main controlling tectonic processes acting on the UK over a million year timescale: regional tectonic stresses and isostatic displacements resulting from ice sheet loading and unloading.

The main regional tectonic stresses acting on the UK crust are ‘ridge-push’ from the North Atlantic mid-ocean ridge and northward indentation of the African plate relative to Europe. The result of these is to give an overall northwest-southeast oriented maximum horizontal stress. Earthquake focal mechanisms are consistent with this orientation. The UK is also subject to the effects of distant plate tectonics and, although much plate-tectonic-related seismicity occurs at depths below 1000 m, it may still be an issue at for a GDF at 200-1000 m depth, through earthquakes and fracture formation/reactivation. The likelihood of fault reactivation of faults intersected by a GDF is very low.

The effect of glacial loading and unloading is potentially of greater significance. Currently the UK is still experiencing isostatic rebound associated with the last glaciation. This results in on-going uplift and seismicity in the north-west of the UK. This isostatic adjustment could affect a GDF by, for example, increased pressure at a glacier base leading to ice melting and “switching back on” of groundwater recharge, which can result in changes associated with faults and fractures at depth. In contrast, in the south-east of the UK, the land-surface appears to be subsiding, possibly as a late stage response to Mesozoic lithospheric stretching beneath the North Sea. Mineralisation, due to mineral-rich fluids flowing through the rock matrix and along the fracture planes themselves, is affected by the hydrogeology and size, distribution and extent of fractures. Mineralisation can effectively lead to the formation of barriers to fluid flow. The impact of fracturing can depend on rock type, for instance, fractures in argillaceous rocks may form and almost immediately heal themselves, whereas fracturing in crystalline rocks may lead to open fractures. Dissolution of earlier fracture or pore fill mineralisation may also occur, leading to gradual changes in groundwater flow paths.

UK SEISMICITY, TECTONIC HISTORY AND VOLCANISM

The UK is characterised by low levels of seismicity and low seismic hazard. However, earthquake activity in the British Isles demonstrates that there is widespread re-activation of pre-existing faults within the Earth’s Crust. The cause and distribution of earthquake activity in the British Isles is complex, but is likely to be dominated by glacio-isostatic deformation or a response to an underlying hot, low-density anomaly in the upper mantle that protrudes south from the Iceland plume. Current evidence indicates that the maximum size of a UK earthquake is unlikely to exceed 6.5 M_w (moment magnitude, or M_w, is a new, universally applicable scale for measuring the severity of an earthquake). The risk to a GDF from direct hazards such as rupture displacement and vibration is very low. Secondary effects such as liquefaction, tsunamis and poroelastic deformation, are considered to be unlikely to affect a GDF, but permanent deformation resulting in permanent changes to porosity and permeability should be considered a potential, but low risk, hazard. Post-glacial isostatic adjustment (GIA) has the potential to increase the maximum magnitude of a UK earthquake to around 7 M_w, which could increase the risk to a GDF following future glacial cycles.

Volcanism is not expected anywhere in the UK for the next few tens of million years.

CLIMATE CHANGE AND GLACIATIONS

Predictions of the extent and timing of future glaciations can be modelled on the basis of proxy evidence of previous glaciations including δO^{18}, amino acid dating, fossil indicator species and geomagnetic polarity records. Consideration of glacial-interglacial cyclicity, driven by 'natural' orbital and anthropogenic CO_2 'forcing', and assessment of the expected duration and degree of influence of anthropogenic warming on this natural cycle, predicts that elevated atmospheric CO_2 will delay the onset of the next glaciation; and no glaciation is expected in the next 170,000 years. The extent of glacial erosion is likely to be in the order of 10's of metres and in occasional circumstances up to around 200 m. Although this is in the range of the shallowest GDF, deep erosion will localise where active ice streams, major glacial melt water drainage routes or major fluvio-glacial outflow incisions occur. Permafrost is known to extend to depths of up to 1000 m under prolonged cold climatic conditions though it is unlikely that permafrost to such a depth has occurred or will occur in future in the UK where the maximum depth is not likely to exceed 500 m. Such permafrost development has the potential to affect the engineering properties of 'soft' rocks, lead to the development of new fracture pathways, affecting groundwater recharge and discharge and affect the engineered elements (e.g. clay and cement-based backfill/buffer materials) of a GDF. Halite is a 'dry' rock so it will not be affected by permafrost, although the components of a GDF built in halite rocks might be affected.

Changes to sea-level could mean that the relative position of a GDF site could be changed so that it is further from or nearer to the coast, or even beneath the sea bed. While the coastal erosion or deposition will be limited to a few 10's of metres, affecting only the surface parts of a GDF. Sea-level changes are likely to alter the groundwater flow paths by changing the base level of discharges.

Global warming induced sea-level rises carry a risk of flooding during the active phases of GDF construction, waste emplacement and closure in low-lying coastal locations. Post-closure flooding of a GDF site will reduce groundwater driving heads and therefore probably reduce groundwater flows in the vicinity of a GDF.

WEATHERING AND EROSION

A review of weathering and erosion processes of the Earth's surface, subsurface weathering and associated processes which affect ground or surface water movement indicates that the greatest depth of weathering or erosion is likely to be caused by glaciation. For example, whereas denudation rates of most hard rock types under non-orogenic conditions are likely to be well below 50 m Myr^{-1} and river incision rates in the UK have been measured up to a maximum of 160 m for the Thames River over the Quaternary (about 65 m M^{yr-1}), over-deepening of glacial troughs may extend to around 200 m, which would affect a shallow GDF if sited in an area of potential over-deepening (i.e. an existing valley). Additionally, subsurface weathering by microbes can influence the migration behaviour of radionuclides through microbially induced pH and redox changes, alteration of mineralogy, consumption/production of radionuclide binding ligands, biofilm formation and gas production. These processes may be affected by increased or decreased supply of nutrients, water, electron donors and acceptors (oxygen), for example, through altered groundwater flows as a result of glacial erosion.

MAIN IMPLICATIONS FOR A GDF

For the majority of processes covered in this review, the possible effects on a GDF are likely to be minimal in all of the geological environments considered over the next one million years. There are a number of processes that may have an effect on a UK GDF, depending on location, but the likelihood of significant consequences is low. In particular the following processes have been identified as having the potential to affect a GDF if circumstances are unfavourable:

- Glacial erosion (relatively near surface only);
- Permafrost;
- Erosion and weathering (relatively near surface only);
- Seismicity (low probability); and,
- Changes groundwater flow patterns.

Until specific sites have been identified as potential locations to host a GDF it is not appropriate to undertake a detailed assessment of many future changes noted above, that may affect a GDF in a generic, non-site specific way. There are a number of issues relating to the formation and thawing of permafrost during cold climatic conditions that would benefit from further generic research. Once potential sites have been identified and rock types are known, specific factors relevant to the site can then be assessed.

Contents

Figures

Tables

Abbreviations

ACLIN	Astronomical Climate Index Model
AFT	Apatite Fission Dating
AGCM	Atmospheric General Circulation Model
ANDRA	Agence Nationale pour la gestion des Déchets RAdioactifs (French agency for nuclear waste)
AOD	Above Ordnance Datum
AP	After Present
BigRAD	BIoGeochemical gradients and RADioactive transport
BIIS	British and Irish Ice Sheet
BIOCLIM	Modelling Sequential BIOsphere Systems Under CLIMate Change for Radioactive Waste Disposal
BIS	British Ice Sheet
BP	Before Present
BVG	Borrowdale Volcanic Group
EA	Environment Agency
EBS	Engineered Barrier Zone
EDZ	Excavation Damage Zone
EQUIP	Evidence from QUaternary Infills for Palaeohydrogeology
ESM	Earth System Models
ESMIC	Earth Systems Models of Intermediate Complexity
ESR	Electron Spin Resonance
FEP	Features, Events and Processes
FIS	Fennoscandian Ice Sheet
GCM	General Circulation Model
GDF	Geological Disposal Facility
GENIE	Grid ENabled Integrated Earth system suite of models
GGF	Great Glen Fault
GIA	Glacial Isostatic Adjustment
GISP2	Greenland Ice Sheet Project 2
GRA	Guidance on the Requirements for Authorisation
GRIP	GReenland Ice core Project
HBF	Highland Boundary Fault
HLW	High Level Waste
ILW	Intermediate Level Waste

IPCC	Intergovernmental Panel on Climate Change
IRD	Ice Raft Debris
LGM	Last Glacial Maximum
LLW	Low Level Waste
MISS	Marine Isotope Sub-Stage
MOC	Meridional Ocean Circulation
MPR	Mid-Pleistocene Revolution
MTZ	Moine Thrust Zone
NGRIP	North GReenland Ice core Project
NIEA	Northern Ireland Environment Agency
ODP	Ocean Drilling Program
OIS	Oxygen Isotope Stages
OIT	Outer Isles Thrust)
PADAMOT	Palaeohydrogeological Data Analysis and Model Testing
PAGEPA	Development and Testing of Models for Climate Impacts on Groundwaters
PGR	Post-Glacial Rebound
PHYMOL	PalaeoHYdrogeological study of the MOL site
PMIP	Palaeoclimate Modelling Intercomparison Project
PMP	Pressure Melting Point
RATE	Radioactivity And The Environment
RSL	Relative Sea Level
SEPA	Scottish Environmental Protection Agency
SMOW	Standard Mean Ocean Water
SRES	Special Report Emissions Scenarios
STZ	Saline Transition Zone
SUF	Southern Uplands Fault
TDS	Total Dissolved Solids
THMC	Thermo-Hydro-Mechanical-Chemical
THMCB	Thermo-Hydro-Mechanical-Chemical and Biological
UMISM	University of Maine Ice Sheet Model
UKCIP	United Kingdom Climate Impacts Programme
UKCS	United Kingdom Continental Shelf
WBF	Welsh Borderland Fault System

1 Introduction

1.1 BACKGROUND

The Geosphere Status Report (NDA, 2010a) outlines NDA-RWMD's current understanding of the role of the geosphere in a multiple barrier concept for geological disposal. It describes how the geosphere contributes to the safety functions of isolation and containment and summarises the evidence supporting this understanding. As noted in the Geosphere Status Report (NDA; 2010a), a significant effort has been spent internationally in identifying the natural processes that may affect the evolution of the geosphere over timescales relevant to the geological disposal of radioactive waste. This includes two Nuclear Energy Agency reviews investigating the stability of the geosphere in both crystalline and argillaceous settings (Nuclear Energy Agency; 2005; 2009). Natural processes that may affect the stability of the geosphere and therefore have an impact on the biosphere have been identified. The principal concerns are:

- Tectonic effects;
- Climate change effects;
- Uplift and subsidence;
- Volcanism; and
- Diagenesis (evolution of the mineralogy of sediments in response to burial and compaction).

This report considers these processes and identifies their key aspects which may impact on a GDF in the UK over the first one million years after closure.

Because this report draws on recent geological history to help quantify many of the above processes a summary of the division of geological time for the Cenozoic era is provided (Table 1). The errors associated with dating the geological boundaries in this table are within the precision of ages given.

Era	Period	Epoch	Approximate Age (Myr.)
Cenozoic	Quaternary	Holocene	0.0117
		Pleistocene	2.588
	Neogene	Pliocene	5.333
		Miocene	23.03
	Palaeogene	Oligocene	33.9
		Eocene	56
		Paleocene	66

Table 1: Geological time for the Cenozoic era.

1.2 REPORT OVERVIEW

Post-closure safety case development has 'traditionally' been focused on the evolution of a GDF system over the first one million years following closure. This is because this is the period when radioactivity is at its greatest and also when a GDF is evolving most quickly as perturbations to the natural system caused by construction and operation dissipate and the waste form and engineered parts of a GDF evolve, for example through corrosion of metals, degradation of organic materials and carbonation of cement based material. Lever et al (2011) have described the key issues affecting the evolution of the geosphere surrounding a GDF over the timescales considered in a post-closure safety case. It considered the implications of thermal, hydrogeological, mechanical, chemical and gas/non-aqueous liquid phase processes for radionuclide migration and how these issues may be considered in a safety case.

Nevertheless over this one million years period the factors that will have the largest effect on radionuclide migration and biosphere impacts are likely to be those related to climate change, including changes in relative sea-level associated with uplift/erosion and tectonics. The NDA-RWMD's current understanding of tectonics, uplift/erosion and climate change and their effect on the geosphere is summarised in Section 4 of their Geosphere Status Report (NDA, 2010a) and this has been used as the starting point for this report. They have identified four sub-topics of particular significance to the UK, which they consider to be 'key' topics. These are:

- Tectonics;
- Uplift, erosion and subsidence;
- Impact of future climate change; and,
- Groundwater movement.

This report considers:

- Tectonic-related uplift (Chapter 2) and subsidence including seismicity (Chapter 3);
- Volcanism (Chapter 3);
- Climate change and glaciations (Chapter 4);
- Weathering and erosion(Chapter 5) ; and,
- Interactions and groundwater flow (Chapter 5).

In each case, the potential impacts of these processes on a GDF in the UK at a depth of between 200 and 1000 m are discussed and an assessment of their possible magnitude undertaken. It is important to be aware that such changes may result from causes that are not necessarily related to geological processes or relate to geological processes that are remote from a Geological Disposal Facility (GDF). Additionally, both climate change and changes to sea-level in near coastal areas may have an impact on groundwater in terms of, for example, chemistry and groundwater flow pathways. Volcanism is only briefly considered in this report because no volcanic activity has occurred in the UK in last 60 Myr (e.g. Pearson et al 1996) and is highly unlikely to occur in the next 10's of million years.

While this report is focused on future natural change it is appropriate to include the effects of anthropogenically induced global warming which may have an impact on the timing and extent of future glaciations. Thus, this has been included in Chapter 4 on Climate Change and Glaciation.

The geological environments considered in this report, that may be potentially suitable for a GDF, are those currently used by NDA-RWMD in their generic safety assessments and are:

- Higher strength rocks to surface (both in lower and higher altitude terrains);
- Higher strength rocks under sedimentary cover (both in lower and higher altitude terrains);
- Lower strength rocks (in lower altitude terrains only); and,
- Bedded evaporites (in lower altitude terrains only).

These geological environments are described in more detail below.

In many cases the processes discussed here are independent of the above geological environments and are therefore considered generically. Where they have a particular impact on specific geological environments this is specifically considered.

1.3 REGULATORY CONTEXT

In the UK, the Radioactive Substances Act 1993 (HMSO, 1993), which was superseded in England and Wales in 2010 by the Environmental Permitting (England and Wales) Regulations 2010 - EPR2010 (HMSO; 2010), provides the legal framework for controlling the management of radioactive wastes in a way that protects the public and the environment. One of the requirements of this act is that authorisation is required for the disposal of radioactive waste. Responsibility for granting such authorisations for England and Wales resides with the Environment Agency (EA) (in Scotland it is with the Scottish Environment Protection Agency (SEPA) and in Northern Ireland with the Northern Ireland Environment Agency (NIEA)).

The EA and NIEA have produced Guidance on the Requirements for Authorisation (GRA) for the deep geological disposal of radioactive waste (Environment Agency, 2009) which sets out the framework within which they regulate geological disposal facilities. Deep geological disposal in Scotland is not currently considered acceptable by the devolved administration.

In the context of this guidance, geological disposal is a long-term management option involving the disposal of radioactive waste in an engineered underground facility, where the geology (rock structure) provides a barrier against escape of radioactivity and where the depth, taken in the particular geological context, substantially protects the waste from disturbances arising at the surface. Such disturbances include those produced by weather and climate change and by human activity. In this context, “depth” could imply horizontal as well as vertical distance – for example, in the case of a disposal facility sited deep within a mountain. A GDF is a facility that meets the requirements for geological disposal. Such a facility could be entirely on land or could be constructed under the seabed but accessed from land.

The RWMD, as the developer of a GDF, will need to demonstrate that it has developed a detailed understanding of the geological setting of the proposed facility in order to for the EA to authorise waste disposal. Part of this understanding will relate to how a GDF is expected to perform over time following closure as the whole GDF system evolves. While the GRA does not specify over what period this understanding should be developed, a period of one million years is frequently used internationally, and previously in the UK, because it is considered an appropriate timescale over which relevant predictions can be made. It is over this timescale that this report focuses.

1.4 NATURE OF A GDF

1.4.1 Current Generic GDF Design

The following is intended to be a brief description of the current NDA-RWMD generic GDF designs. It is based on the 2010 NDA Geological Disposal Geosphere Status Report (NDA, 2010a) and the 2010 NDA Geological Disposal Near-Field Evolution Status Report (NDA, 2010c). A GDF is a purpose built facility excavated deep within a stable geologic environment, defined for the UK of between 200m and 1000m deep (Department for Environment, Food and Rural Affairs et al 2008). The geological disposal concept is a multi-barrier system that includes the waste form, the engineered facility, backfill and barriers and the geology surrounding a GDF in providing isolation and containment of waste without future intervention. The NDA Geosphere Status Report (2010a) states, 'This involves designing engineered barriers that will work together and in combination with the natural barrier afforded by the geosphere to prevent radionuclides being released to the surface environment in amounts that could cause harm to life and the environment'.

Until sites are identified and host rock geologies known, no specific GDF designs have been prepared. This report does not consider specific GDF designs but considers the case of a generic GDF. Nor are specific wasteform or buffer/backfill systems considered because the majority of impacts resulting from future changes are likely to affect most disposal options.

A GDF will be constructed to dispose of Intermediate Level Waste (ILW) and High Level Waste (HLW) and some Low Level Waste (LLW). It will also accommodate spent nuclear fuel and may include separated plutonium and uranium if the Government decides to dispose of these materials. Most wastes will be conditioned, for example vitrified or grouted into disposal containers prior to disposal. The current Managing Radioactive Waste Safely (MRWS) White Paper Baseline Inventory says that the packaged volume will be 476,000 m^3 and the total radioactivity will be 87,200,200 Terabequerels (TBq).

1.5 POTENTIAL GEOLOGICAL SETTINGS

NDA (2010a) provides descriptions of three geological settings in England and Wales that may potentially be suitable for a GDF. The descriptions are summarized here.

1.5.1 Higher Strength Rocks

NDA (2010a) defines higher strength rocks as '*typically comprising crystalline igneous, metamorphic rocks or geologically older sedimentary rocks, where any fluid movement is predominantly through divisions in the rock, often referred to as discontinuities. Granite is a good example of a rock that would fall in this category*'. The geology of such terrains is complex, as they contain a range of rocks formed from igneous and metamorphic processes. Additionally, they often record many phases of such activity, which have served to rework and deform earlier rocks. The rocks have undergone several phases of 'brittle' deformation, leading to the formation of fractures and faults. Higher strength rocks can also be formed from geologically older sedimentary rocks. These rocks have generally been deeply buried and have become indurated as a result of chemical and physical changes such as the expulsion of water from the sediment and the precipitation of secondary minerals (‘cements’) from circulating groundwaters.

A common feature of all of these rock types is that they contain fractures. Many, in particular crystalline rocks, also have low total porosity (voids in the rock which will generally be filled with groundwater).

1.5.2 Lower Strength Rocks (often referred to as argillageous rock)

NDA (2010a) defines lower strength sedimentary rocks as *‘typically comprising geologically younger sedimentary rocks where any fluid movement is predominantly through the interconnected pore structure of the rock mass itself. Many types of clay are good examples of this category of rocks’*. In France, where a Callovo-Oxfordian clay layer is currently being considered by ANDRA (the French waste management organisation) as a host rock for a GDF, the sequence consists predominantly of layers of clay, sandstone and limestone.

Particularly in the case of marine deposits, argillaceous rocks can form extensive layers and have consistent properties over large areas of many hundreds of square kilometres. Consequently, the typical geometry of a sedimentary rock is a sheet-like body where the horizontal extent is much greater than the thickness. For example, the approximately 150 m thick Callovo-Oxfordian clay layer currently being considered by ANDRA has been shown, through detailed geophysical and geological investigations, to be continuous and of uniform thickness over a 350 km^2 area.

It is important to note that, although sedimentary rocks may have originally been laid down to form laterally extensive bodies, subsequent geological processes may have resulted in them being partially eroded or faulted so that they now form a number of physically distinct if not separate bodies.

In the context of suitability for a GDF, it is necessary that the host rock horizon is sufficiently thick to accommodate the engineered facility. It is also necessary to understand how the rocks above (‘cover rocks’) and below the host horizon affect its containment properties. This is of particular relevance to lower strength sedimentary rocks because of their sheet-like geometry, which means that other rocks may be present in close proximity to a GDF.

1.5.3 Evaporites

NDA (2010a) defines evaporites as *‘typically comprising anhydrite (anhydrous calcium sulphate), halite (rock salt) or other minerals that result from the evaporation of water from water bodies containing dissolved salts’*. The minerals are very soluble, and the persistence of evaporites over geological time demonstrates that there has been negligible groundwater flow in the formations in which they are found. Evaporites typically form in cyclic sequences, in which minerals precipitate in a well-defined sequence controlled by their solubility and the extent to which the original body of water has been concentrated by evaporation. Evaporites, particularly those formed in marine environments, can form over extensive areas and attain large thicknesses. For example, Permian deposits in the Delaware Basin in the USA extend over 300,000 km^2 whilst in Texas salt deposits up to 3,500 m thick have been found. The Salado Formation, in which the Waste Isolation Pilot Plant (WIPP) in New Mexico is constructed, is a massive bedded halite formation approximately 530 to 610 m thick.

Original evaporite sequences are stratified (that is, they form horizontal ‘beds’) but because they are less dense than most other sedimentary rocks and are capable of plastic deformation, they can displace the overlying sediments and rise upwards in inverted tear shaped structures known as diapirs. Such structures are common in evaporite environments; for example, the German HLW radioactive waste disposal programme is considering salt diapirs as potential host rocks.

There are no known salt diapirs below the UK mainland although extensive bedded evaporites are present on-shore and in the near shore continental shelf. Thick bedded rock salt (halite) deposits in the UK occur in rocks of Permian and Triassic age. Mercia Mudstone Group, of Triassic age, provides 90% of the UK's resource of rock salt; halite formations have maximum known thicknesses of approximately 400 m.

1.6 POST CLOSURE EVOLUTION OF A GDF

Although a site for a UK GDF has not yet been selected and the geological setting and host rocks are therefore not yet determined, it is possible to give a general summary of the various time frames of significance for the post-closure evolution of a GDF over the first one million years and the role of the geosphere in these time frames (from NDA 2010a). Note that the following time frames are at the order of magnitude level and are not absolute. They will be dependent on a number of factors including waste types, waste form and geological environment:

- **The first thousand years** - In the earliest post-closure years, the hazard from the waste would be at its highest. After about a thousand years, depending on the specific composition of the waste, the shorter-lived radionuclides would have largely decayed and the radionuclide inventory would have significantly reduced. The safety objective of a GDF during this time period is to contain the waste. During this period, the principal safety function of the geosphere is to isolate the waste and protect the Engineered Barrier System (EBS). It achieves this by providing a stable geological setting with several hundred metres of rock between a GDF and the surface;
- **A thousand to of the order of a hundred thousand years** - During this period, there is likely to be substantial breakdown of components of the engineered system and small amounts of radionuclides may be released into the geosphere. Therefore, understanding the evolution of the geosphere and its stability over this timescale becomes particularly important. The geosphere would continue to provide isolation for a GDF and, in conjunction with the buffer and backfill, provides a degree of containment for radionuclides released from the waste packages, although it will not necessarily contain poorly sorbed, long-lived radionuclides;
- **A hundred thousand to about a million years** - Some components of the engineered system may have degraded substantially whilst other components, such as clay-based barriers, may retain their original properties. The geosphere continues to provide isolation for a GDF and, through the processes of geochemical retardation and immobilisation, acts to contain the small proportion of radioactivity that has migrated from a GDF. Only the longest-lived radionuclides would persist into this time period. The hazard from the residual radioactivity in the EBS would resemble that from a natural uranium ore body.

Following closure of a GDF, groundwater flow paths should return more or less to their pre-construction configuration rapidly, in the case of fractured 'hard' rocks over a few decades, though this will be dependent on a number of factors. It is likely that oxidising conditions developed during the construction and operation of a GDF that will have affected the surrounding rocks will also return to a reducing state fairly rapidly. However some of the changes within the rock mass induced by oxidation will be irreversible or will only revert to pre-construction conditions over very long time scales.

Given the mid continental plate location of the UK it is reasonable to assume that the current tectonic regime will continue. As such the UK will continue to be subject to relatively modest earthquakes, mainly associated with recognised fault zones. Without a new glaciation, isostatic adjustment of the land surface following the last glaciations will continue at a slowing rate as equilibrium is regained, with southern Britain sinking and northern Britain rising relative to sea-level. Unless directly affected by fault reactivation caused by an earthquake, neither of these processes are expected to have any effect on a GDF at depth.

If over the next one million years if climatic conditions remain similar to the UK's current temperate climate, which is unlikely, then average weathering and erosion rates will be modest, of the order of a few 10's of metres (see Chapter 5) and are not expected to have any effect on a GDF at depth. However, in some upland and adjacent areas deeper erosion, perhaps up to a maximum of 200 m, may occur.

If the current global warming trend continues, even at a slowed rate, sea-level will rise as water currently held in ice caps and glaciers melts and thermal expansion of sea water occurs. This will lead to submersion of low lying coastal areas except in areas where isostatic rebound equals or exceeds sea-level change.

Most models, even taking into consideration anthropogenic global warming, suggest that the northern hemisphere will return to glacial conditions (see Chapter 4) in the future. This situation is likely to occur several times over the next one million years. Even if the site of a GDF is not actually glaciated it will probably experience permafrost conditions.

In glaciated conditions, sub-glacial and outwash streams may create deep channels perhaps eroded up to 100 m into the existing land surface. Oxygen rich sub-glacial meltwater may recharge groundwater to depths not normally experienced under temperate conditions. While such groundwater fluxes may be buffered by the rock mass (with respect to oxygen) relatively quickly it may be expected that they will have some effect at depth.

If permafrost conditions prevail, depending on temperatures and rock types, frozen ground conditions may be experienced to depths of 100's, perhaps 1000, metres. Under such conditions the near surface environment (perhaps to a few 10's of metres) may experience freeze/thaw conditions causing cryoturbation of near surface materials and rapid breakdown of rocks. At depths where permanent freezing is present, groundwater flows will be disrupted or prevented and gases may be trapped in or near to a GDF. Such conditions may lead to changes in groundwater chemistry.

1.7 GROUNDWATER

Groundwater movement is the most important process that eventually leads to radionuclide migration from a GDF (except in evaporites). Groundwater gets into a GDF system by infiltrating though the rock mass and the engineered barriers. Groundwater will be present within the rocks around a GDF during and after its construction, which may require the excavation to be pumped during construction and operation. Once pumping stops on sealing and closure of a GDF, groundwater levels that were depressed by pumping will gradually return to the pre-construction natural level. A clay or salt hosted GDF is likely to be 'dry' with no or minimal groundwater inflows during operation.

Groundwater movement through a GDF system will depend on the hydraulic conductivity of the system and its development over time. This will be a function of the performance of the engineered barriers through time. Water movement through a GDF system is likely to be by diffusion through barriers or by advection through imperfections or failures in the engineered barriers and backfill. Depending on the host rock type groundwater movement through the host rocks will be largely by advection in the case of fractured ‘hard’ rocks or diffusion in the case of mudrocks and by a combination of both for rocks with intermediate properties. In the case of salt there will be almost no groundwater flow.

A distinctive feature of periods when land is covered by sea is recharge of heavy, saline marine water into the system. The location and orientation of fractures will play an important role in determining the trajectories of saline water in displacing freshwater masses.

Solutes will also be present within the groundwater as a result of the interaction between the water and the rock (including minerals) through which it flows. Solutes may also be present in the groundwater around a GDF as a result of the movement of the interface between saline and fresh water or the exchange of solutes between saline and fresh water likely to be caused by changes in hydrogeology during cycles of glaciation and deglaciation and related sea level change.

Solute movement will occur through a combination of diffusion through barriers within the system and diffusion or advection within groundwater. This will be dependent on the solute and hydraulic gradients within a GDF. Solute gradients will be a function of differences in solute concentrations in water inside a GDF and the development of material diffusive characteristics through time. If the materials inside parts of a GDF have very low water contents, the diffusion gradient will not operate as effectively as would be the case when all parts are saturated or near saturated and so solute transfer into such low water content regions is likely to be small or negligible.

1.8 SUMMARY OF KEY CONSIDERATIONS

1.8.1 Uplift and subsidence

There are two main controlling processes acting on the million year timescale, which are regional tectonic stresses and isostatic displacements due to ice sheet loading or unloading, causing vertical displacements of the Earth’s surface. Chapter 2 summarises current knowledge of the mechanics of tectonic uplift and subsidence and associated erosion and burial that is pertinent to understanding the UK setting in relation to a GDF sited in the UK with a focus on the next million years.

The main tectonic stresses acting on the UK crust are ridge-push from the North Atlantic mid-ocean ridge and northward indentation of the African plate relative to Europe. The result of these is to give an overall north-west to south-east oriented maximum horizontal stress. Earthquake focal mechanisms are consistent with this.

Perhaps more significant on the million year timescale is the effect of glacial loading and unloading. Currently the UK is still experiencing an isostatic rebound because of the melting of the ice sheets associated with the last glaciation. This is associated with ongoing uplift and seismicity in the north-west of the UK (Bradley et al; 2009). In contrast, in the south-east of the UK, the land-surface appears to be subsiding, possibly as a late stage response to Mesozoic lithospheric stretching beneath the North Sea.

1.8.2 Tectonics

A review of the current understanding of past and present tectonics relevant to the various specified geological settings in the UK, and how these may influence the evolution of the geosphere and thus impact the biosphere of those geological settings is given in Chapter 3. This includes current seismicity and the existing stress regime, and looks at the structural geology and tectonic evolution of the various geological settings in the UK, with a view to understanding the potential for current and future tectonic activity and fault reactivation. Neotectonic fault activity has also been considered.

The British Isles has been subject to multiple episodes of deformation and, unlike plate boundaries where stress regimes are generally straight-forward, the driving forces of present-day deformation are less obvious. Although there are relatively strong variations in the spatial distribution of earthquake activity throughout the British Isles, observed seismicity cuts through the major geological structural boundaries, most of which run north-east to south-west. This means it is difficult to explain earthquake activity in terms of these structures. The situation is further compounded by earthquake activity due to human activities, such as coal and mineral extraction. Clearly, ongoing earthquake activity in the British Isles demonstrates that there is widespread re-activation of pre-existing faults within the Earth’s crust. The orientation of the present day stress field, to some extent controls, which faults are most likely to be activated in the 1 million years time frame considered in this report. However, the small size of the earthquakes makes it difficult to accurately map earthquakes to specific faults, particularly at depth, where fault distributions and orientations are unclear. Of particular importance to the postclosure safety case for a GDF is the fact that, no British earthquake recorded either historically or instrumentally has produced a surface rupture.

1.8.3 Climate change/glaciation

Future changes in climate have the potential to produce the largest changes on a GDF system, not only directly controlling global eustatic sea-level, but also by governing the rate and form of erosion and sediment dispersal from the land surface. In particular, the future timing, intensity and frequency of glacial and interglacial episodes will be fundamental in determining the long term evolution of the biosphere and geosphere both regionally and globally. This glacial-interglacial cyclicity will directly influence rates and styles of erosion across the UK landmass, from those that are fluvial-dominated during temperate periods, to glacial erosion, dominated by both topographically controlled glaciers and by topographically enveloping ice sheets, during glacial episodes. The nature, thickness and extent of ice cover, as well as duration of each glaciation will control amounts of isostatic depression of the land surface, associated with ice sheet loading, and uplift due to rebound, which will control the relative sea-level and ground water regime changes across the UK.

Chapter 4 is a review of current published research and modelling of the nature of future climate change, concentrating on the projected frequency and duration of glacial and interglacial episodes and assessing their intensity relative to those documented from the Quaternary record (in particular those from the last 500 ka). It considers modelled glacial-interglacial cyclicity, driven by ‘natural’ orbital and anthropogenic CO_2 ‘forcing’, and assesses the expected duration and degree of influence of anthropogenic warming on this natural cycle. Of particular importance in the context of this context is the low likelihood of UK glaciation within 100 kyr.

1.8.4 Diagenesis

Diagenesis normally applies to sedimentary rock systems but the geochemical and rock-water interaction processes involved may be similar in crystalline or fractured rock systems. It is broadly defined as the cumulation of chemical, physical or biological changes taking place in a sediment between its deposition and the completion of lithification or cementation, and before the onset of metamorphism (Frey, 1987). Diagenetic processes are often slow and occur over long timescales, and some diagenetic processes, traditionally considered important in the evolution of aquifers, hydrocarbon reservoirs and caprocks may not significantly affect the geosphere or biosphere over the 1 million year time frame under consideration in the context of radioactive waste disposal.

There are a number of processes that will influence diagenesis, including changes in sea-level and climate change that may be important during the timescales of GDF relevance. In particular, changes to groundwater flow paths and hydrochemistry during glacial and permafrost phases could introduce 'weathering' processes to greater depths than are prevalent during temperate climatic phases with alteration of, for example, fracture wall rock mineralogy that in turn may change the properties of parts of the geosphere, with consequent impact on the biosphere.

Diagenesis can be divided into Eodiagenetic (early), Mesodiagenetic (burial) and Telodiagnetic (late) processes. There is of course a continuum between these diagenetic "stages". Mineralogical and geochemical changes associated with these classes will be of varying significance.

1.8.5 Palaeohydrogeology

Palaeohydrogeology has been defined as, *"a combination of observations on hydrochemical and isotopic differences in various groundwater zones or bodies, mineralogical data on the rock formations and the hydraulic properties of the same formations, which are compiled to allow interpretation of the evolution of the rock-water system over long time intervals in the past.*" (NEA, 1993; Bath et al., 2000a).

Essentially, palaeohydrological studies aim to understand past conditions that have determined the movements and compositions of groundwaters. The relevance and application to radioactive waste management is based on the concept that understanding the impact of past drivers on the groundwater system may enable the potential impact of similar future drivers to be predicted.

The time intervals of interest encompassed in palaeohydrogeology can range anywhere between the short timescales seeking to understand the impact of anthropogenic influences and the very long timescales of geological processes. A one million year timescale can be considered to be "hydrogeological", rather than "geological" because it corresponds with the inferred age range of the groundwaters that occupy many of the systems of interest (e.g. PALAEUX, 1999; Bath et al., 2000a and b; Marivoet et al., 2000; Boulton et al., 2001). In this respect, observation of telodiagenetic processes may provide valuable information that will inform palaeohydrogeological investigations.

2 Tectonic related uplift and subsidence

2.1 INTRODUCTION

The Earth is a dynamic planet with all points on the land surface (and the rocks that lie beneath it) being subjected to processes that cause vertical movement and which result in either uplift or subsidence (McKinley and Chapman, 2009). These vertical movements of Earth's surface can vary considerably in magnitude and occur over a wide range of areas and timescales in response to both tectonic and non-tectonic processes (e.g. glacial unloading, igneous intrusion and/or volcanic processes). Upward vertical movement (uplift) forms topography, which generally results in erosion. Downward vertical movement (subsidence) creates accommodation space, which generally results in burial. Over long (geological) timescales erosion and burial may further enhance uplift and subsidence, respectively and it is therefore important to understand the fundamental drivers of uplift and subsidence (e.g. tectonic or non-tectonic) and the relationship with erosion and burial (Litchfield et al, 2009).

Additionally, the spatiotemporal behaviour of earthquakes within continental plate interiors is different from that at plate boundaries (Calais et al., 2010). Fennoscandia and northern Europe, including the United Kingdom Continental Shelf area has experienced geologically recent (post glaciations) large magnitude earthquakes of up to M_w5 (e.g. Davenport and Ringrose, 1985; Davenport et al., 1989; Ringrose et al., 1991; Fenton, 1992; Stewart et al., 2000; Lagerbäck and Sundh, 2008). Earthquake rupture zones at plate margins are rapidly re-stressed by continuing tectonic motions, making the location of recent earthquakes and the average time between them consistent with the faults' geological, palaeoseismic and seismic histories and thus relatively predictable. In contrast, large earthquakes within continents and plates are episodic, often clustered and appear to migrate with time. Whilst such fault behaviour is recognised in many continental interiors, what determines the activation of a particular mid-continental fault and controls the duration of its seismic activity remains poorly understood (Crone et al, 2003).

Tectonic uplift occurs over time periods of many hundreds of thousands to millions of years, which is generally too long for the time frame of one million years being considered in this report. However, as will be seen below, significant uplift over shorter periods of several tens to hundreds of thousand years occurs through processes of deglaciation, which may be linked to and driven by tectonic processes.

In the case of a GDF, unless there is rapid erosion, uplift over a broad geographical extent is unlikely to have a direct influence on GDF performance, because it is anticipated that the mechanical and hydrogeological state of the deep host rock would be little perturbed by this process (McKinley and Chapman, 2009). An exception could be where uplift results from isostatic rebound following ice loading, with relatively rapid (~ mm yr^{-1}) uplift associated with changing sea-level there may be significant impacts on GDFs sited near the coast or offshore beneath the sea and accessed from land (McKinley and Alexander, 2009; McKinley and Chapman, 2009).

The main causes of vertical movements and their timescales relate to (McKinley and Chapman, 2009):

- Short-term cyclic motions, caused by planetary orbits, for example, minute daily movements due to gravitational effects (Earth's tides);

- Larger, millennial-scale effects of glacial loading or unloading and sea-level variations caused by climate changes driven by Milankovitch variations in the Earth's movement around the Sun, or other mechanisms, and;

- Longer timescale, tectonic processes resulting from plate-motion-driven crustal deformation and the emplacement and evolution of magma bodies, which result in more significant and longer term movements.

Uplift and erosion of the uplifted surface tightly linked processes, with erosion rates dependent on the mechanical and chemical properties of the rocks, climate, altitude and uplift rate (see Chapter 5); high uplift rates generally correlate with high erosion rates. Equally, subsidence is generally accompanied by the creation of accommodation space (often basinal structures) and sedimentation onto the sinking surface, with erosion to the flanks of the subsiding area (McKinley and Chapman, 2009).

For nuclear facilities and the storage of waste, the impacts of uplift and erosion are generally more significant than subsidence and burial (McKinley and Chapman, 2009; Litchfield et al., 2009). This chapter, therefore, summarises current knowledge of the mechanics of tectonic uplift and subsidence and associated erosion and burial that is pertinent to understanding the UK setting in relation to a GDF sited in the UK with a focus on the next million years.

The Atlantic continental margins have long been considered as the archetypal 'passive continental margin', so called because continental crust and oceanic crust are part of the same plate and the presumed relative tectonic quiescence follows a syn-rift phase of faulting and magmatic activity during continental breakup (e.g. Bond and Kominz, 1988; Stoker and Shannon, 2005). However, the basins on the margins bordering Norway, the UK, the Faroes and Ireland and those intracratonic basins inboard and adjacent to passive margins contain mild and occasionally strong compressional structures of Cenozoic age, which have been the subject of increasing debate. The timing, magnitude, nature and causes of this uplift have remained uncertain and in part controversial, but these factors are now becoming more fully understood. However, it is becoming increasingly apparent that there is a common history in the Neogene (Miocene to Holocene) development of the Atlantic margin of north-west Europe between Mid-Norway and south-west Ireland (e.g. Swiecicki et al., 1998; Doré et al., 1999; Roberts et al., 1999; Brekke, 2000; Stoker et al., 2002, 2005a) and that tectonic events have been ongoing across the region, overlapping in time from area to area.

The impact of mantle- and lithospheric-scale processes affecting intraplate areas has only recently been recognized in coastal realms (Cloetingh, 2000; Cloetingh et al., 2003; Cloetingh et al., 2005). The present state and behaviour of the Earth's System is a consequence of processes operating on a wide range of time scales. This includes the long-term effects of tectonic uplift and subsidence on river systems (Holbrook and Schumm, 1999), residual effects of the ice ages on crustal movement, natural climate and environmental changes over the last millennia, and also the powerful anthropogenic impacts of the last century. Understanding the present state of the Earth's System to predict its future for man to engineer use of it, requires that this spectrum of processes, operating concurrently but on different time scales, needs to be better understood. The challenge is to describe the state of the system, to monitor its changes, to forecast its evolution and, in collaboration with others, to evaluate modes of its sustainable use by the human society (Cloetingh et al., 2005).

The UK and its continental shelf areas currently lie in an intracratonic setting within the north-west European plate, an area extending from the Alpine Mountain chain in the south, northwards across the Alpine Foreland to the ('passive') north-east Atlantic Continental Margin and North Atlantic spreading axis to the northwest: an area often referred to as the north-west 'Alpine' Foreland (e.g. Williams et al, 2005). Within this region, strata are distributed in a series of failed rifts and extensional sedimentary basins many of which developed as a result of episodic rifting associated with the progressive northwards propagation of a spreading axis associated with the opening of the North Atlantic in Mesozoic and Cenozoic ('Tertiary') times (e.g. Roberts, 1975, 1989; Ziegler, 1982). The basins and UK landmass area thus generally lie adjacent to and inboard from the north-east Atlantic margin in a tract that runs from Mid-Norway to southwest Ireland and which suffered similar deformation episodes related to:

- The opening of the Atlantic Ocean to the north-west as a spreading axis propagated northwards from late Cretaceous times, and;
- Compressional events associated with the closure of the Tethyan Ocean far to the south.

However, although surrounded largely by basins containing structures that reflect multiple episodes of broad regional extensional (subsidence) and compressional events throughout Phanerozoic times, including thick sequences of late Neogene (Pliocene-Holocene) sediments offshore, the present day outcrop geology and distribution of geological units of the British Isles and surrounding continental shelf areas demonstrate that they are themselves elevated and show uplift. Offshore, regional angular unconformities in late Neogene successions attest to major changes in sedimentary styles and uplift of areas to supply large quantities of sediment (e.g. Stoker, 2002; Stoker et al., 2005a; Sejrup et al., 2005a).

2.2 PROCESSES LEADING TO TECTONIC RELATED UPLIFT AND SUBSIDENCE

2.2.1 Mechanics of tectonic uplift and subsidence

With obvious connections to isostacy, tectonic uplift and subsidence cause major changes in the elevation of the Earth's surface, with most uplift and subsidence the result of tectonic processes that either thicken, thin or flex the Earth's crust and which occur at differing spatial locations and timescales (Litchfield et al., 2009):

- Thickening and thinning (extension) of the crust – leads to changes in the density structure of the crust resulting in vertical movements due to isostatic responses, and;
- Flexure of the crust – driven largely by horizontal compression or loading, but also during extension, which may also lead to vertical movements due to isostacy.

The tectonic processes may be on regional (tectonic plate) scale, driven by mantle and asthenospheric processes and resulting in both mountain building and deep oceanic basin formation or local scales (few kilometres), producing differential uplift and subsidence generally seen across fault blocks and folds. Other causes of uplift and subsidence are possible, including both tectonic and non-tectonic processes (e.g. volcanic and glacial processes).

The main drivers and processes causing these changes in surface height (uplift/subsidence and tectonic or non-tectonic) and the relationship(s) with erosion and burial include:

- Tectonic subsidence – crustal thinning (extension) and tectonic loading/crustal flexuring:
    - Tectonic loading and crustal/lithospheric flexure;
    - Foreland deformation – foreland basins (foredeeps) and forebulges;
    - Local scale tectonic subsidence;
    - Continental/lithospheric extension – isostacy and geometrical constraints (extensional faulting);
    - Continental/lithospheric extension - pure-shear or simple-shear lithospheric extension, and;
    - Pure-shear or simple-shear lithospheric extension;
- Anomalous tectonic subsidence;
- Tectonic uplift:
    - Orogenic movement or uplift (orogenesis or orogeny);
    - Wrench (strike-slip) faulting, and;
    - Epeirogeny (uplift or subsidence of the crust without folding), epeirogenic movement and inversion tectonics, or sedimentary basin inversion.

2.2.2 Rates of movement or deformation

The following briefly outlines the rates of various tectonic processes. They are generalisations, but provide limits to assist in the assessment of the potential for any particular process to affect a GDF site in the time frame considered. As indicated earlier, tectonic processes probably occur over periods that are too long for the one million years being considered here.

In general, horizontal tectonic movement at late boundaries is in the range 1-10 cm yr^{-1}, with the current fastest relative plate movement being up to 15 cm yr^{-1} at the East Pacific rise (Cloos, 2009). Movements of >5 cm yr^{-}1 are classified as fast, giving rise over 1 Myr to displacements of 50 km and over 100 kyr, to displacements of circa 5 km. It is now recognised that most earth movements are localised at plate boundaries and generally in the range 4 to <1 cm yr^{-1} (Kreemer et al., 2000). As no plate boundaries are close to the UK, this is considered to be of negligible risk for a UK GDF.

2.2.3 Present-day UK stress field

Discussions of potential uplift mechanisms, either epeirogenic or tectonic basin inversion, should consider the stress fields affecting the north-west European Shelf margin, including both historical and the current day situations. In southern and central England and the southern North Sea the maximum horizontal (compressive) stress is oriented roughly northwest to north-northwest, with the minimum horizontal stress oriented perpendicular to this, northeast to east-northeast. Northwards the stress field may veer gradually, such that the maximum horizontal stress direction may become north-south in central Scotland (though rather poorly-constrained hereabouts) before swinging round to roughly east-west in the northern North Sea. In the central North Sea stress data are markedly confused.

On a larger scale, the UK lies within a regional zone of broadly consistent northwest-southeast horizontal compressive stress orientations stretching from the Alps to northern Britain. This arises from plate movements and forces associated with the east to south-easterly separation of Europe from North America and the roughly northerly movement of the African plate relative to Europe (Brereton and Muller 1991).

Despite the overall compressive nature of the stress field, the lack of visible deformation in post-Miocene strata (younger than c. 10 Myr) preserved in the North Sea basins indicates that tectonic sedimentary basin inversion is not presently occurring. This suggests that for the last 10 million years or so, the UK area has been subject to a low stress regime a consequence principally of the ongoing process of `ridge-push' with an absence of `high stress' episodes because of Alpine continental 'docking'.

The absence of strong compressive or tensile stresses is borne out by *in situ* stress measurements from boreholes (Figure 1) which indicate that at depths greater than about 500 m, the vertical (overburden) stress is intermediate (σ_2) between two orthogonal horizontal stresses. Such a stress regime is compatible with strike-slip rather than reverse or normal faulting (Figure 1).

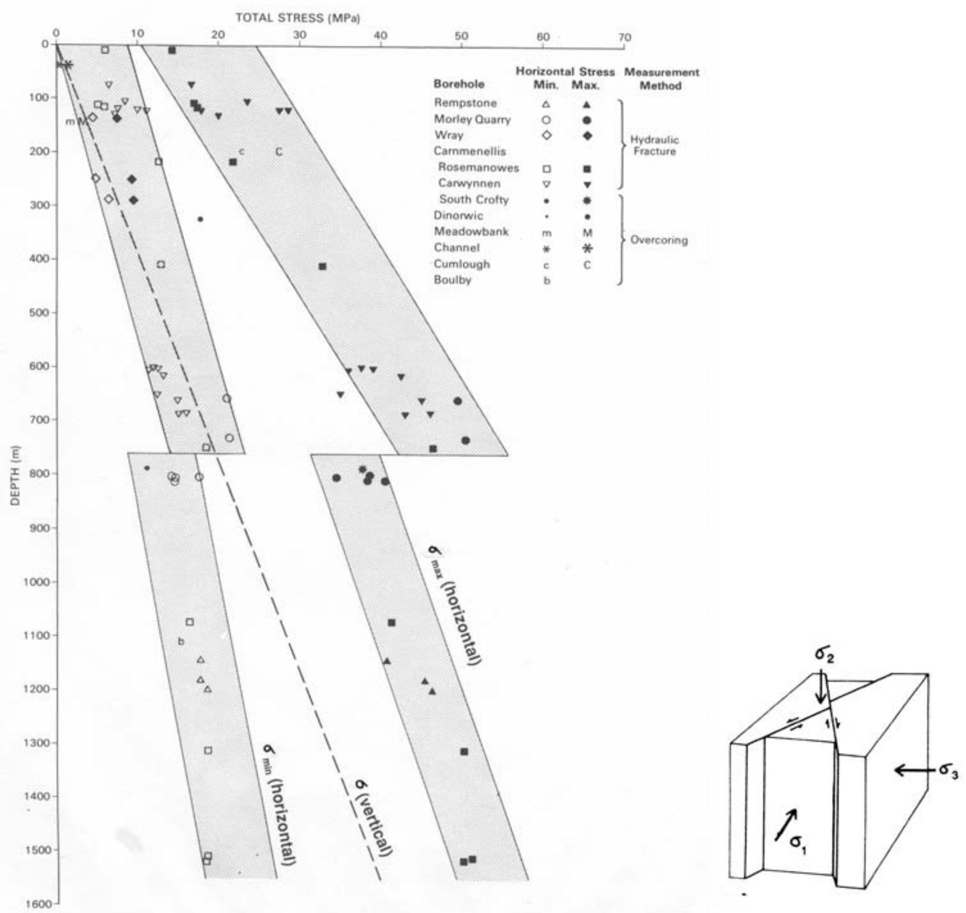


Figure 1. Crustal stress data from selected UK boreholes (adapted from Evans, 1987). Note that at depths >600 m σ2 is vertical i.e. strike slip regime.

Focal mechanisms for recent UK earthquakes (Figure 2) are consistent with the observed stress field. They all show a component of strike-slip motion, with two of the events almost pure strike-slip and another three dominantly strike-slip. Of the remainder, three events show dominantly reverse movements, though with subsidiary strike-slip. Another point of note is the fact that for all but one of these events the maximum horizontal stress (horizontal component of σ_1) lies in the north-west to south-east quadrant, and the minimum horizontal stress (horizontal component of σ_3) lies in the north-east to south-west quadrant, in excellent agreement with the stress map (Figure 2).

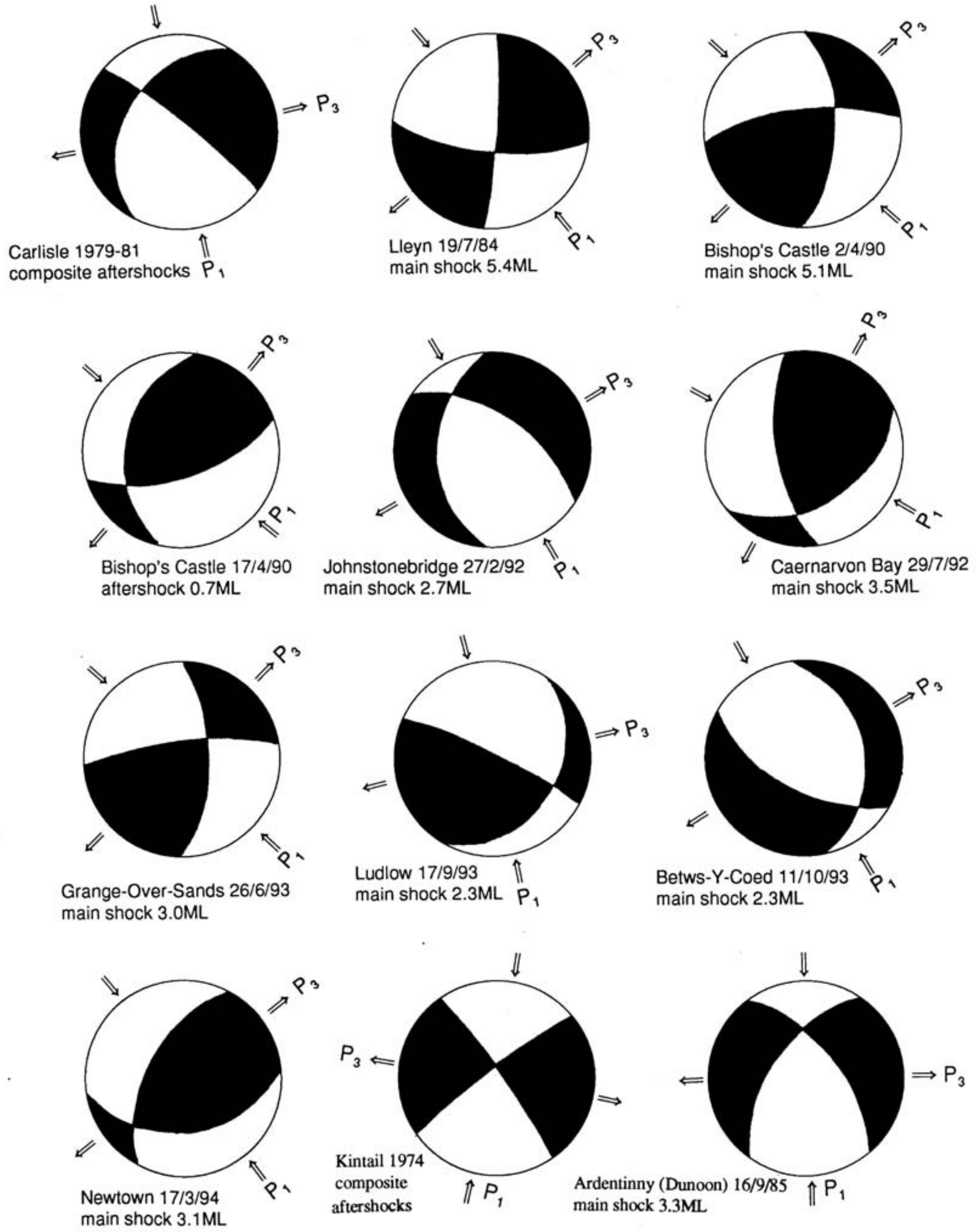


Figure 2. Summary focal mechanisms for selected UK earthquakes, plotted on upper focal hemisphere. Compressional quadrant shown in black, dilational quadrant shown in white. Also shown are the maximum (σ) and minimum (σ3) stresses projected on to the horizontal plane. Data abstracted from Assumpcao (1981), Marrow and Roberts (1984), Turbitt et al. (1985), Redmayne and Musson (1987), Ritchie et al. (1990) Walker (1993, 1994, 1995).

2.2.4 Magmatic Underplating

Rapid uplift is caused by a process known as magmatic underplating. Underplating is more common close to continental margins. When continental crust passes over mantle anomalies rising magma from partial melting may be trapped beneath or within the lower continental crust leading to thickening of the crust which in turn leads to uplift. Once an area has passed over a mantle anomaly crustal subsidence may then occur. Northern England, Southern Scotland and, probably, Northern Ireland have experienced underplating during the Palaeogene (from about 65.5 to 23 Myr ago) with up to 3km of uplift as a result which has subsequently been eroded (Tiley et al; 2004). The erosion of this thickness of strata had largely occurred by the close of the Palaeogene.

Underplating can cause rapid uplift and erosion of thick sequences of strata over relatively short geological timescales (10’s of Myr). The last underplating event to affect the UK ended over 20 Myr ago and the UK is not currently subject to this process.

2.3 NEOTECTONICS: PRESENT DAY STRESS FIELDS AND NEOTECTONIC STRUCTURES, GLACIAL ISOSTATIC MOVEMENTS (REBOUND), GLACIO-SEISMOTECTONICS AND SEISMICITY

The effects of glacial loading and isostatic rebound are increasingly recognised as providing significant uplift (in the order of several hundred metres to kilometre scales) in some regions of the North Atlantic margin and adjacent continental hinterland.

Neotectonics is a sub-discipline of tectonics that considers aspects of structural geology, seismology, regional geophysics, geomorphology and engineering geology (Hancock and Williams, 1986). It is the study of current (or recent, in geological time) motions of the Earth’s crust, particularly those produced by earthquakes, and is aimed at understanding the geological and geomorphological processes that lead to motion and deformation of the Earth’s crust, particularly those produced by earthquakes, with the goals of understanding the physics of earthquake recurrence and seismic hazard embodied in these processes. The corresponding time frame is referred to as the neotectonic period, although there has been disagreement as to how far back in time "geologically recent" is. Whilst some suggest that the neotectonic period extends from the middle Miocene, neotectonics is commonly viewed as the youngest, still unfinished stage in Earth tectonics (Pliocene – Recent) and for some is essentially synonymous with "active tectonics". More general agreement is emerging that the actual time frame may be individual for each geological environment, but extended sufficiently far back in time to fully understand the current tectonic activity (Dramis and Tondi, 2005). The term may also refer to the motions/deformations in question themselves where it is used to indicate a suite of structures that were initiated and propagated in a late Cenozoic tectonic stress field that has persisted with little or no change of orientation until the present day and usually when differential stresses are less than those necessary for active faulting or the development of other mesoscopic structures (Hancock and Engelder, 1989; Hancock, 1991).

Until recently, research on neotectonics and related continental topography development has, in the main, focused on active plate boundaries characterized by generally high deformation rates (refer Cloetingh et al., 2005). The intraplate sedimentary basins and rifts of the Northern Alpine foreland and north-west European Shelf margins are associated with a much higher level of neotectonic activity than hitherto believed (Figure 3). Seismicity and stress indicator data, combined with geodetic and geomorphological observations, demonstrate that Europe’s intraplate lithosphere is being actively deformed (Cloetingh et al., 2005). This has major implications for the assessment of its natural hazards and environmental degradation. Neotectonics plays a significant role in the evolution of the geomorphology and soils of an area, determining areas of active sedimentation, pedogenesis and erosion (in upland regions). It leads to tilting and sagging of large blocks and faulting that can result in for example shifting and influencing the courses of large rivers, creating increased sinuosity and changes to local climates.

Recent high spatial resolution in the quantification of earthquake hypocentres (Figure 3) and focal mechanisms, and vertical motions of the land surface seismicity studies and geomorphological evidence from Brittany (Bonnet et al., 1998, 2000), Normandy and the Channel and Dover Straits areas (Lagarde et al., 2000; Van Vliet-Lanoë et al., 2000, 2002), southern England (Preece et al., 1990), the Ardennes–Eifel region (Demoulin et al., 1995; Demoulin, 1998; Meyer and Stets, 1998; Van Balen et al., 2000), the Upper Rhine Graben (Nivière and Winter, 2000) and the North German Basin (Ludwig, 1995; Bayer et al., 1999) demonstrate the important contribution of neotectonics to the topographic evolution of intraplate Europe (Cloetingh et al., 2005). For example, in intraplate areas such as the Pannonian Basin (Gábris, 1994) and the Rhine–Meuse delta (Berendsen, 1998; Stouthamer and Berendsen, 2000; Cohen et al., 2002), the importance of Late-Quaternary neotectonic activity on fluvial deposition has been recognized.

This section, therefore, reviews the developments in understanding crustal deformation and seismicity during glaciations, and following deglaciation, glacio-isostatic rebound and the potential contributions to uplift and neotectonic movements and deformation.

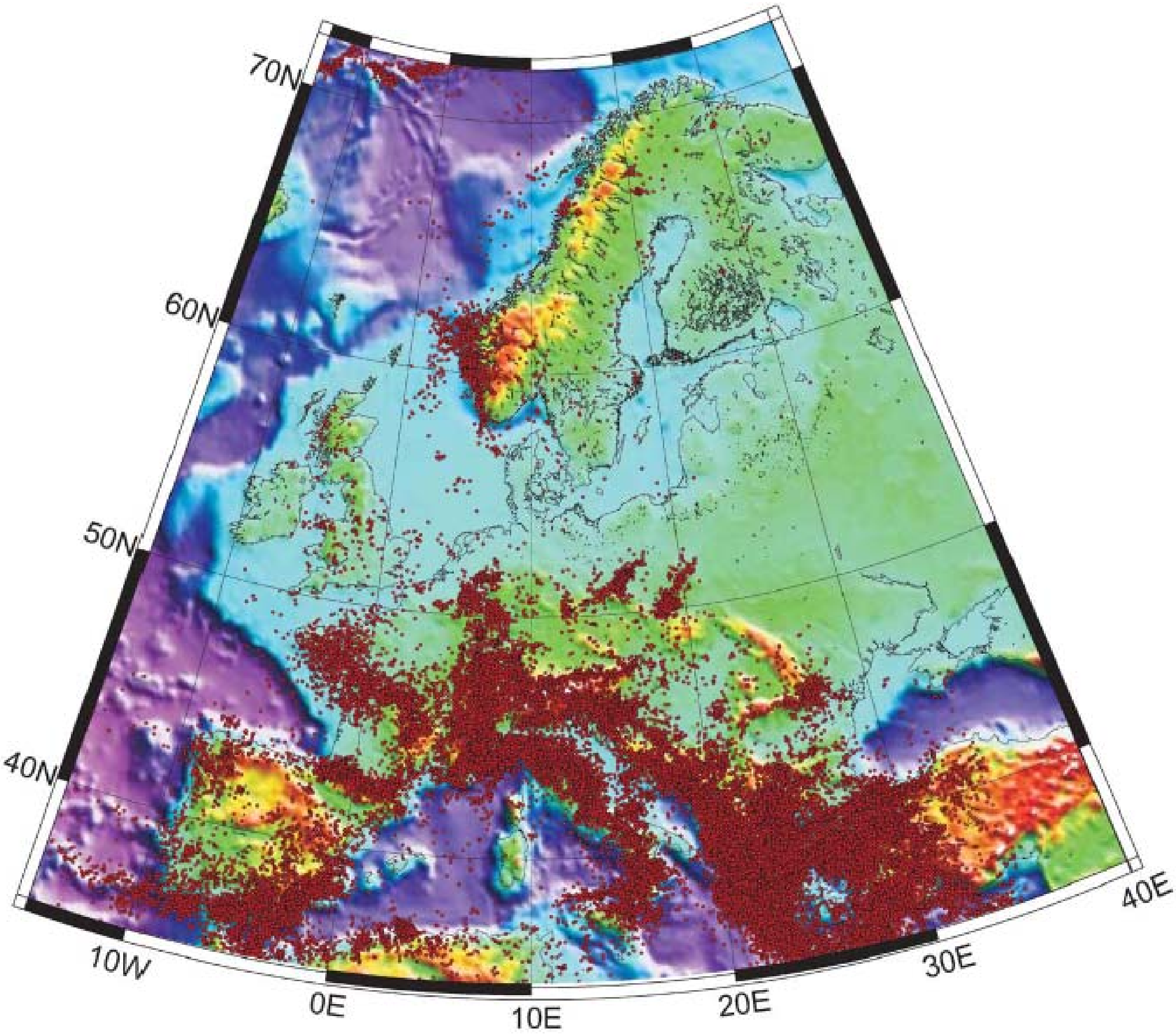


Figure 3. Seismicity map of Europe, illustrating present-day active intraplate deformation (from Cloetingh et al., 2005). Background elevation images are extracted from the ETOPO2 data set. Earthquake epicentres are from the NEIC data centre, and are shown as red dots. Reprinted from Cloetingh, S., Ziegler, P.A., Beekman, F., Andriessen, P.A.M., Matenco, L., Bada, G., Garcia-Castellanos, D., Hardebol, N., Dèzes, P. and Sokoutis, D. 2005. Lithospheric memory, state of stress and rheology: neotectonic controls on Europe's intraplate continental topography. Quaternary Science Reviews, 24, 241-304 with permission from Elsevier.

2.3.1 Neotectonic structures in southern England and France as indicators of late Neogene regional crustal stresses

The term 'joint' is used to describe a fracture, that is not a stylolite or vein, on which, in the field, it is barren and where there is no evidence for offset related to shear, dilation or pressure solution (Hancock, 1985). Neotectonic joint systems are cracks representing the most recent and last systematic brittle structures to form within a region subject to uplift and erosion (Hancock and Engelder, 1989). Although related to regional stress fields, the initiation and propagation of neotectonic joints occurred close to the Earth's surface, generally within the upper 0.5 km of the crust. These shallow joint systems generally form within the upper 0.5 km of the crust because unloading as a result of denudation and lateral relief consequent on uplift are prerequisites for their propagation and they are not generally found in cores taken from greater depths (Engelder, 1985). They occur throughout areas of at least 10,000 km^2 and where *in situ* stress measurements and fault plane solutions are available (*e.g* Engelder, 1982, 1985; Bevan and Hancock, 1986; Hancock and Engelder, 1989), these joints are shown to be parallel or sub-parallel (less than <20°) to the direction of maximum horizontal stress of the contemporary stress field.

In southern England and northern France, a regional system of generally northwest-trending late Cenozoic mesofractures cut Upper Cretaceous and Palaeogene rocks (Bevan and Hancock, 1986; Hancock, 1991). Although a regionally developed system of northwest-trending mesofractures had not been recognized previously, a 130°-striking set of vertical joints was identified by Toynton (1983) in the chalk of northern Norfolk, and Middlemiss (1983) described 120-130° striking sets of vertical and steeply inclined joints cutting the chalk exposed on the east coast of Kent. However, neither author had interpreted the fracture sets in terms of mechanical classes or causative stress fields.

The northwest-trending system cross-cuts an additional, earlier, east-west trending fracture system associated with similarly trending flexures formed during the Oligocene to Early Miocene 'Helvetic' phase of Alpine deformation. The northwest-trending fractures and lineaments thus indicate northwest-directed compression and east-west extension which could be related to far field effects of either Atlantic ridge-push forces, or Alpine compression. Bevan and Hancock (1986) point out that the neotectonic joints are parallel to neotectonic normal faults in the Lower Rhine embayment, which become less well developed to the west with increasing distance from these major structures. The Lower Rhine faults were reactivated and propagated into the Quaternary cover as a consequence of north-east-south-west regional tension generated during the late Neogene to Recent 'Jura' phase of northwest-southeast Alpine convergence. The northwest-trending neotectonic extensional structures in southern England and northern France are of about the same age and related to the same stress regime as the neotectonic normal faults of the Lower Rhine embayment (Bevan and Hancock, 1986). However, because fractures of the northwest-system cutting Palaeogene sediments in southeast England are truncated by the erosion surface beneath the Plio-Pleistocene Red Crag deposits, it is possible that in England fracture initiation did not continue into Quaternary times.

The Alpine orogeny occurred when the continents of Africa and India and the small Cimmerian plate collided (from the south) with Eurasia in the north. Convergent movements between the tectonic plates (the Indian plate and the African plate from the south, the Eurasian plate from the north, and many smaller plates and microplates) had begun in the early Cretaceous, but the major phases of mountain building began in Palaeocene to Eocene times, with folds in the Wessex-Channel Basin formed during Miocene uplift (e.g. Chadwick, 1986, 1993). Currently the process still continues in some of the Alpine mountain ranges.

Vertical or inclined planar fissures that may for the most part be classified as joints are described from exposures of both the Coralline and Red Crag in East Anglia (Balson and Humphries, 1986). Measurements of fissure orientation reveal orthogonal patterns of alignment in both formations. Balson and Humphries (1986) suggest that the field relations of the fissures to the host Crag sediment suggest a tectonic rather than periglacial origin and suggest that the fissure system is the product of early Pleistocene tectonic flexuring in the area, on the western margin of the subsiding southern North Sea basin. Commenting on their open and filled nature, Hancock (1991) suggested that they have a different appearance to and are unlike the fracture systems in the Chalk of southern England and France. These Chalk fracture systems are closed and terminate upwards in the London Clay at the unconformity beneath the Red Crag (Bevan and Hancock, 1986) and are therefore likely to be of differing origin.

2.3.2 Post glacial rebound or glacial isostatic uplift, neotectonics and implications for induced faulting and GDFs

Understanding the causes, location and potential magnitude of glacioisostatic rebound and glacially-induced faulting is of direct relevance to radioactive waste management programmes because these periods are so long that the effects of ice sheet conditions and glacially-induced faulting need to be assessed regardless of inferred global climate warming, because the global warming effect is likely to taper off before the end of the assessment period (e.g. Solomon, 2007; SKB, 2006b; Lund and Näsland, 2009).

It is difficult to separate the effects of regional uplift and exhumation arising from late Neogene tectonic and mantle related processes from those of glacial isostacy and glacioseismicity. Indeed, it seems that there is an increasing likelihood of a continuum between processes. Results of geothermochronological studies have provided constraints on the timing and magnitude of uplift of the North Atlantic borderlands, demonstrating a polyphase record that commences as early as in Oligocene times, and that is interpreted as the combined result of upper mantle perturbations, intraplate stresses and post-glacial rebound (refer Hendriks et al., 2005).

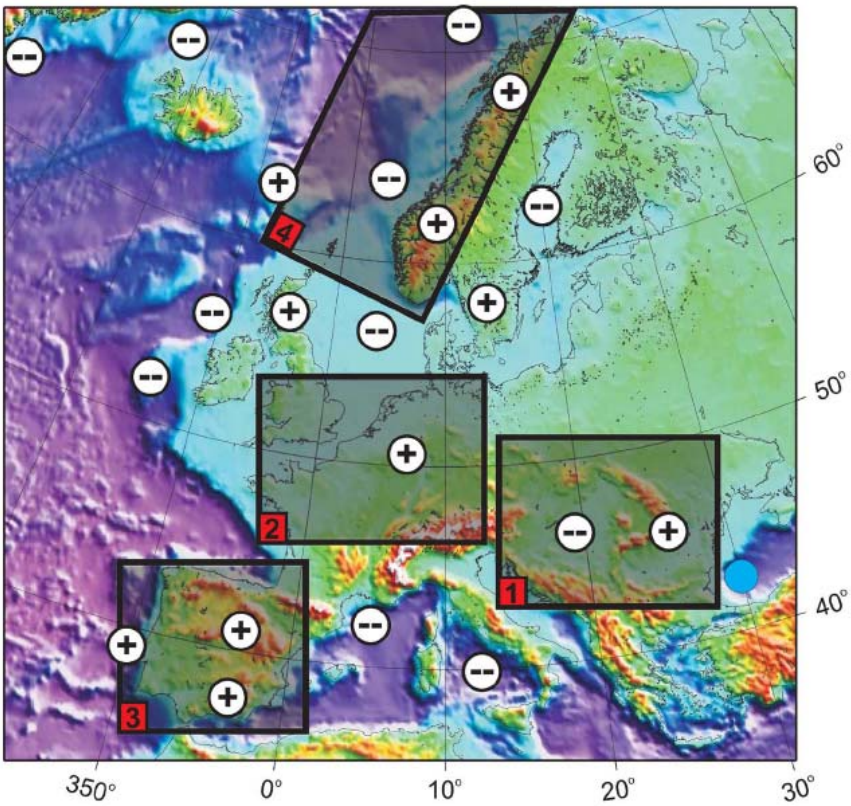


Figure 4. Topographic map of Europe (from Cloetingh et al. 2005), showing intraplate areas of Late-Neogene uplift (circles with plus symbols) and subsidence (circles with minus symbols). The areas indicated are those studied by Cloetingh et al. (2005) when investigating neotectonic controls on Europe's intraplate continental topography: (1) the Carpathian–Pannonian Basin System; (2) northwestern European platform; (3) microcontinent Iberia; (4) northern Atlantic continental margin. Reprinted from Cloetingh, S., Ziegler, P.A., Beekman, F., Andriessen, P.A.M., Matenco, L., Bada, G., Garcia-Castellanos, D., Hardebol, N., Dèzes,

P. and Sokoutis, D. 2005. Lithospheric memory, state of stress and rheology: neotectonic controls on Europe’s intraplate continental topography. Quaternary Science Reviews, 24, 241-304 with permission from Elsevier.

There is evidence of late Neogene uplift across Europe and the north-east Atlantic margins (Figure 4; Cloetingh et al., 2005.) and following deglaciation of large areas, that this is still ongoing and related to isostatic uplift. Furthermore, earthquake studies show that contemporary seismicity in the Scottish Highlands is concentrated in and around the glacio-isostatic uplift centre, and it is said to be clustered in areas of Late Quaternary shoreline dislocations, faults and palaeoseismic phenomena (Davenport et al., 1989; Musson, 1996). From these and other related studies, there is therefore, an emerging picture of some form of linkage between recent and on-going crustal movements in Scotland and the region's recent glacial history. Many of the above studies (e.g. Firth et al., 1993; Davenport et al., 1989; Musson, 1996) have speculated on the extent and nature of the coupling between seismotectonic and glacial processes. However, it remains unclear whether the post-glacial tectonics of the Scottish Highlands are driven by the far-field tectonic stresses imposed by mid-Atlantic ridge push and Alpine collision or near-field effects of glacio-isostatic rebound, or both (Firth and Stewart, 2000). There is a growing recognition that glaciation and plate tectonics have combined to shape the Earth over geological time (Eyles, 1993), and that the redistribution of mass that occurs during the growth and decay of ice sheets may perturb the Earth's rotational state (Stewart et al., 2000).

The concept of glacio-isostasy, the idea that major glacial advances and retreats are accompanied by significant crustal deformation, was proposed more than a century ago (Jamieson, 1865, 1882; De Geer, 1888; Kolderup, 1913). Strong theoretical and empirical support now exists for the theory that major glacial advances and retreats, due to the growth and decay of continental ice sheets over the span of many centuries to several millennia or tens of thousands of years, are an important agent of dynamic change within the solid Earth. The formation and subsequent melting of large ice masses (sheets, glaciers *etc*.) places sudden, geologically speaking, variations in stress loads on the lithosphere. The glacial loading and unloading effects are deep seated and involve both elastic deformation (flexing) of the Earth’s crust and lithosphere and viscous flow in the mantle, (Figure 5) which as a consequence, affect large areas (Pascal et al., 2010). During glaciations global sea-levels fall as water is locked up in the ice masses. Conversely, during deglaciation, sea-levels rise as meltwaters return to the oceans: the size attained by ice sheets at the last glacial maximum meant that global sea-level fell by about 120 m and therefore, during the same glaciation phase, the global lowering of sea-level may also contribute to eustatic unloading and lead to the exposure of some areas of the continental shelves (Stewart et al., 2000). The on-going viscous response of the solid Earth to the melting of the last great Late Pleistocene ice sheets are currently regarded as global in extent and spanning long time periods, being a significant contributor globally to changes in crustal levels (Figure 6) and intraplate deformation. Glacial isostacy is now recognised as driving changes in sea-level and land elevation, and importantly, causes readjustment of lithospheric stresses in formerly glaciated areas and adjacent regions, where increased seismicity and powerful earthquakes are documented.

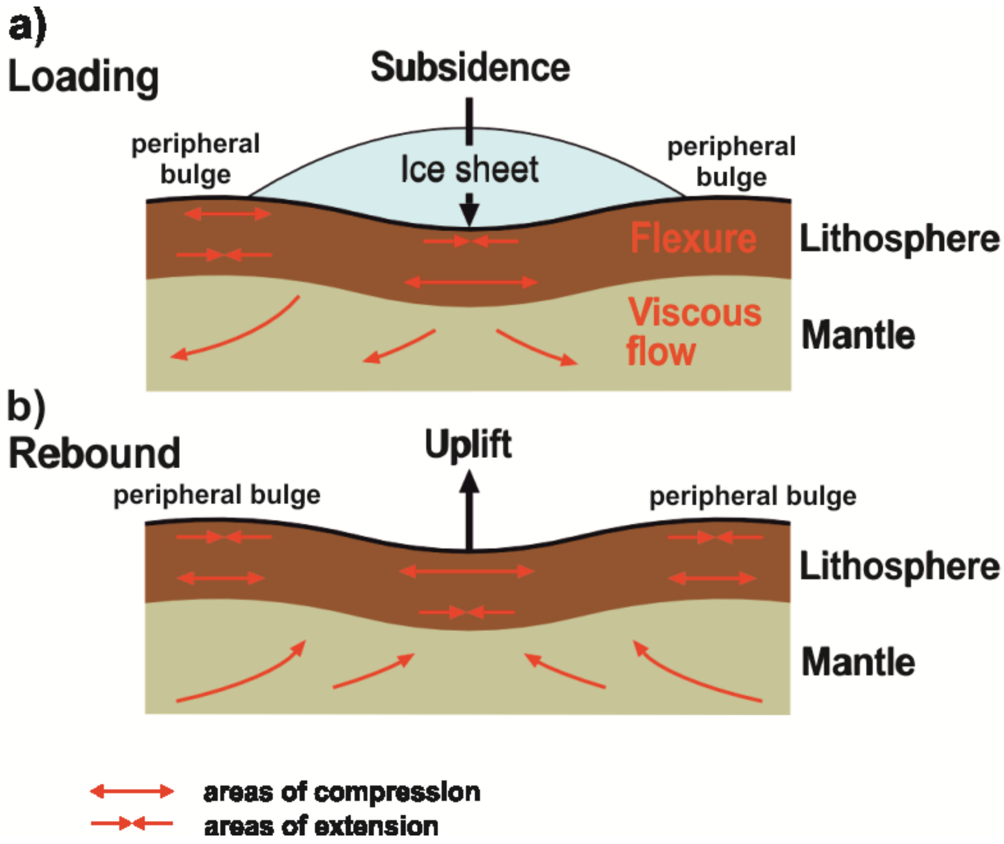


Figure 5. Schematic illustration of the loading and unloading effects on the crust, lithosphere and mantle, with the development of a peripheral bulge during the emplacement and subsequent decay of an ice sheet.

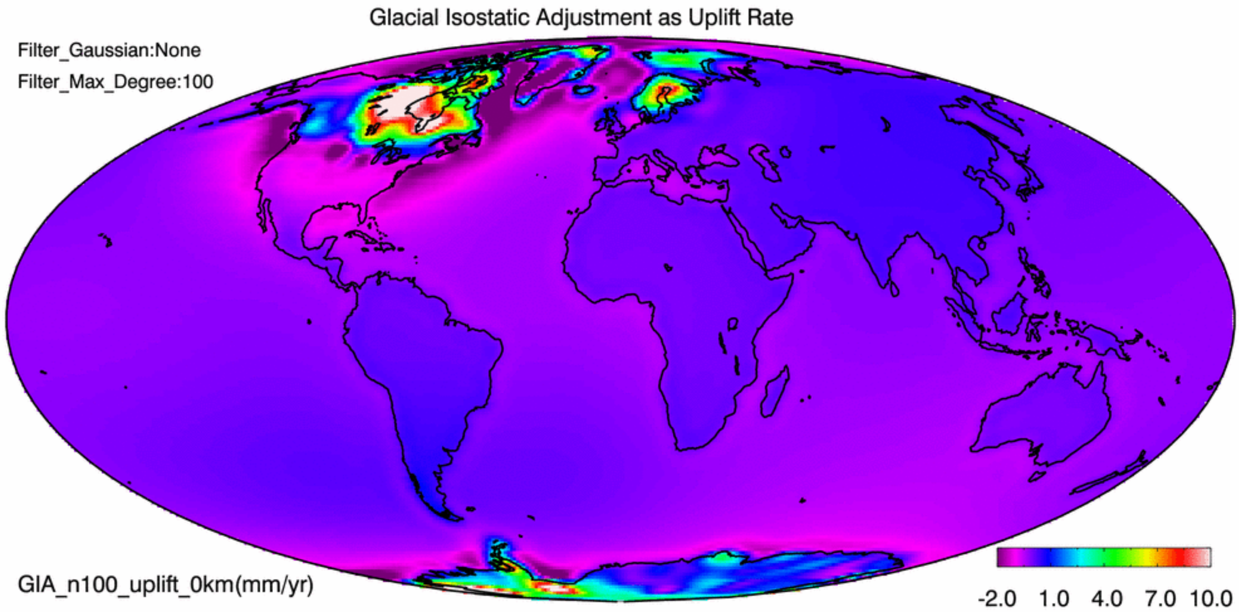


Figure 6. Rate of lithospheric uplift due to Post-glacial Rebound. Image from "http://grace.jpl.nasa.gov (Wahr and Zhong, 2013)

Evidence exists for glacially induced fault movements having occurred as ice sheets retreated at locations in northern Fennoscandia (Lagerbäck, 1979; Olsen et al., 1998) and in continental North America (e.g. Kenner and Segall, 2000). Rapid uplift of land due to glacial rebound is taking place in Fennoscandia where studies have indicated that decimetre scale shear movement has occurred on fractures, which could potentially jeopardise spent fuel disposal canister integrity (SKB, 2006a). Assessments of safety at GDFs must take into consideration the possibility of faulting occurring. A deep GDF must be located to avoid intersecting a large fracture zone with the potential for a large magnitude earthquake – not because of the shaking, but more the possibility of spent nuclear fuel canisters might be compromised (Chapman et al., 2009; Lund and Näsland, 2009).

The melting of large ice caps over the span of many centuries to several millennia is, therefore, an important agent of dynamic change within the solid Earth. It induces significant redistributions of mass on the surface of the Earth, drives sea-level and land elevation changes, and readjusts lithospheric stresses in formerly glaciated areas and adjacent regions. Effects manifest as increased seismicity and, eventually, in powerful earthquakes: the discovery of the 'endglacial' faults of northern Scandinavia (Kujansuu, 1964; Lagerbäck, 1979; Olesen, 1988) proved definitively that glacial loads and, therefore, climate can act as tectonic drivers over very short geological times (Pascal et al., 2010).

However, it remains unclear whether the post-glacial tectonics of, for example, the Scottish Highlands and other areas of northern Britain, are driven by the 'far-field' tectonic stresses resulting from mid-Atlantic ridge-push and Alpine collision, or 'near-field' effects in previously glaciated regions of glacio-isostatic rebound, or both (e.g. Firth and Stewart, 2000; Bradley et al., 2009). Emerging glacial rebound models support the view that to explain the distribution and style of contemporary earthquake activity in former glaciated shields of eastern Canada and Fennoscandia, both tectonic stresses and glacial rebound stresses are required.

2.3.2.1 GLACIAL ISOSTATIC ADJUSTMENT

During the last glacial maximum about 20,000 years ago, much of Asia, North America, Greenland, Antarctica, northern Europe and the central parts of the land mass of Great Britain, were covered by ice sheets up to three kilometres thick (Boulton, 1985) and radii approaching 1000 km in Fennoscandia and 2000 km in Laurentia. Ice sheets in Iceland, the British Isles and the mountain chains of Europe were smaller and thinner (Stewart et al., 2000). As this ice melted, so the loading on the land mass altered, the result of which is the gradual and long-term re-adjustment of the land mass, such that the north of Great Britain is undergoing uplift whilst the south is sinking (see below).

Redistribution of mass occurs during the growth and decay of ice sheets over timescales of several tens of thousands of years. The term glacial isostatic adjustment (GIA) is now more commonly used with the recognition that the response of the Earth to glacial loading and unloading is not limited simply to downward land movement and upward rebound movement. It involves horizontal crustal motion (Johansson et al., 2002; Sella et al., 2007), changes in global sea-levels (Peltier, 1998), the Earth's gravity field (Mitrovica and Peltier, 1993; Mitrovica et al., 2001), induced earthquakes (Wu and Johnston, 2000; Stewart et al., 2000; Firth and Stewart, 2000) and changes in the Earth's rotational motion (Wu and Peltier, 1984; Mitrovica et al., 2001).

Glacial loading

During periods of glaciation, great thicknesses of ice may accumulate over large continental areas, placing enormous weight on the Earth's crust: 1 km of ice generates a pressure of *c.* 90 bars (Pierce et al., 2002). During the last glacial maximum *c.* 20 ka, the Laurentide ice sheet of Canada and the United States and the Fennoscandian ice sheet had maximum thicknesses of 2.5 to 3 km (Lund and Näsland, 2009). The ice sheet over Great Britain during the last glacial maximum is thought to have attained a maximum over Central and north-west Scotland (Boulton et al., 1991; Firth and Stewart, 2000), achieving a thicknesses of *c.* 1 km. These ice sheets formed slowly and, as in the process of tectonic loading, their weight resulted in the Earth's crust being gradually deformed and down warped with the viscoplastic mantle material forced to flow laterally away from the loaded region to make room for the flexing crust (Figure 5, Figure 7 and Figure 8).

As described by Stewart et al. (2000), prior to glaciation, crustal deformation proceeds at a rate and in a style that is driven largely by horizontal plate motions, but with the onset of glaciation the prevailing tectonic stress state (field) is overprinted by a glacial stress (Figure 7). As ice sheets build-up and wane, so they exert fluctuating glacial loads, which result in a lagged crustal response because whilst the elastic response to unloading/loading is instantaneous, the viscoelastic response of the mantle is slower. Additionally, the rheology of the mantle and the density contrast between ice and rock ensures that the magnitude of the crustal adjustment is less than the imposed ice mass fluctuations, with the crustal depression curve being of lower amplitude and not 'showing' the rapid fluctuations of ice thickness Figure 8). Elastic crustal flexure results in a 'bowl' of depression below the ice sheet centre, accommodated by crustal contraction as deeper mantle material flows radially outward from below the maximum ice load (Figure 7). Horizontal crustal motions induced by the ongoing glacial isostatic adjustment increase out from the ice sheet centre and are greatest at the ice sheet margins (James and Bent, 1994; Peltier, 1998). In the upper crust, flexural stresses imposed by an ice load of a few km thickness will be a few tens of MPa (Walcott, 1970; Stein et al., 1989),which may typically match or exceed the magnitude of regional tectonic stresses (*e.g.* the present-day maximum horizontal compression (σ_H) in eastern North America is ~10 MPa (Adams, 1989b)).

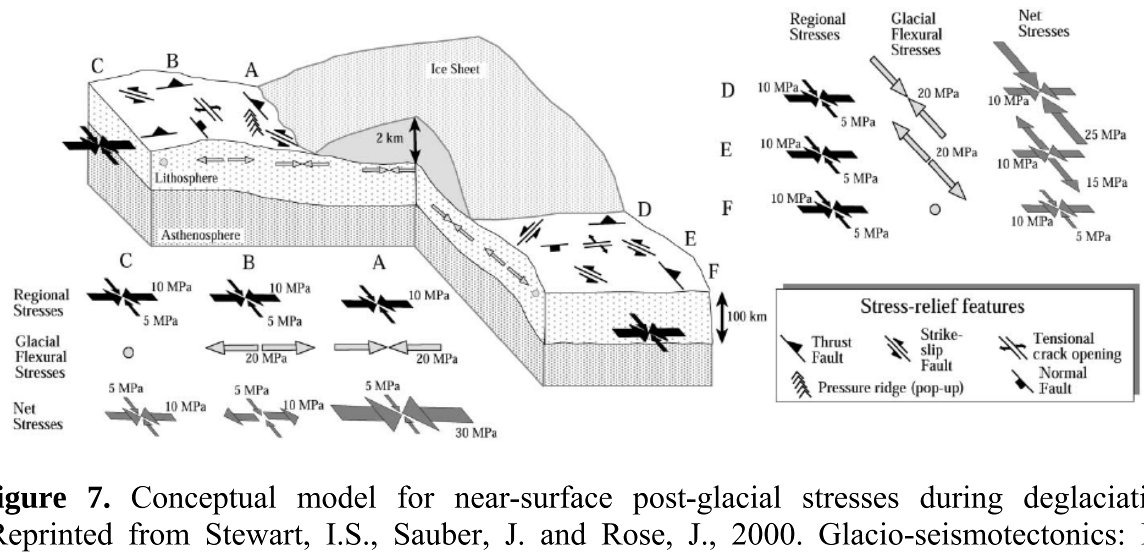


Figure 7. Conceptual model for near-surface post-glacial stresses during deglaciation (Reprinted from Stewart, I.S., Sauber, J. and Rose, J., 2000. Glacio-seismotectonics: Ice sheets, crustal deformation and seismicity, Quat. Sci. Rev., 19, pp. 1367-1389 with permission from Elsevier). The exploded block model illustrates how the superimposition of changing glacial flexural stresses (light grey arrows) on a uniform regional tectonic stress field (black arrows) results in contrasting resultant stress states (dark grey) and different styles and orientations of stress-relief phenomena at the ice margin (A and D), the forebulge (B and E) and the undeformed foreland (C and F). Modified from Adams (1989a) after Walcott (1970). Note the considerably exaggerated vertical scale difference between the lithosphere thickness and the ice sheet thickness.

The ice sheet imposes a load that has an effect beyond the ice sheet margin, out into the ice-free foreland. Upper crustal flexural upwarping occurs which when combined with deep-seated, inward flow of mantle material exuded from beneath the ice mass, creates uplift and radial crustal extension within an uplifted forebulge, or peripheral bulge (Figure 7 and Figure 8; *e.g.* Mörner, 1977; Fjeldskaar, 1994; Lambeck, 1995; Stewart et al., 2000). The forebulge may extend for several hundreds of kilometres beyond the ice margin (Lund and Näsland, 2009). Land beneath the thickest ice is pushed down by up to half a kilometre, whilst uplift of the forebulge and land areas beyond the ice sheet was less, being in the order of a few tens to a maximum of several hundred metres. Extensional stresses of up to 10 MPa are predicted for a forebulge related to a roughly 1 km thick ice load (Walcott, 1970; Stein et al., 1989). An additional consideration in the formation of the forebulge is the elasticity of the lithosphere, which may give rise to downwarping of the crust, not only under the ice sheet but also outside the ice margin. This gives rise to a flexural depression, in which lakes from glacial meltwater may form between the ice margin and the forebulge.

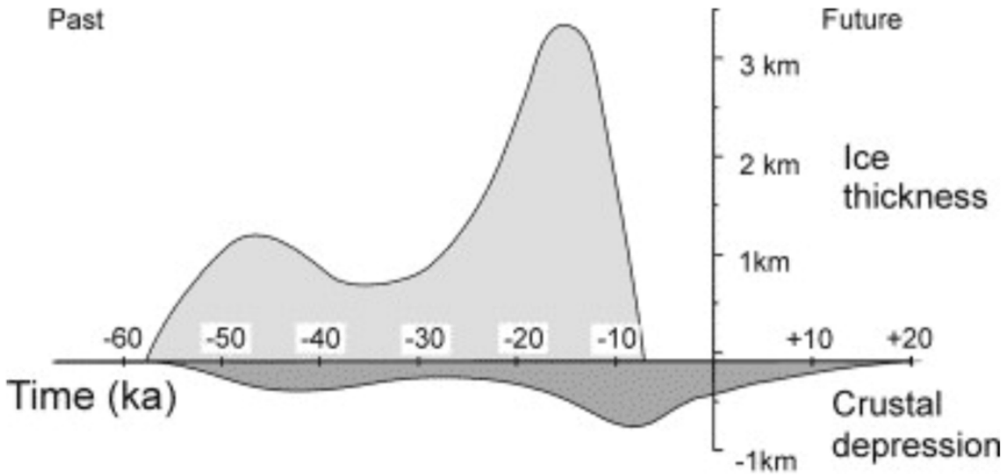


Figure 8. Simplified loading history for central Fennoscandia during the last (Weichselian) major glaciation (Reprinted from Stewart, I.S., Sauber, J. and Rose, J., 2000. Glacio-seismotectonics: Ice sheets, crustal deformation and seismicity, Quat. Sci. Rev., 19, pp. 1367-1389 with permission from Elsevier). Ice thickness appears above the time axis and consequent crustal depression appears below. The model illustrates how rapid vertical ice loading and unloading changes lead to far slower crustal depression or rebound, and that the scale of crustal response is considerably less than the imposed ice-mass fluctuations, reflecting viscous flow of the mantle (Pascal et al., 2010).

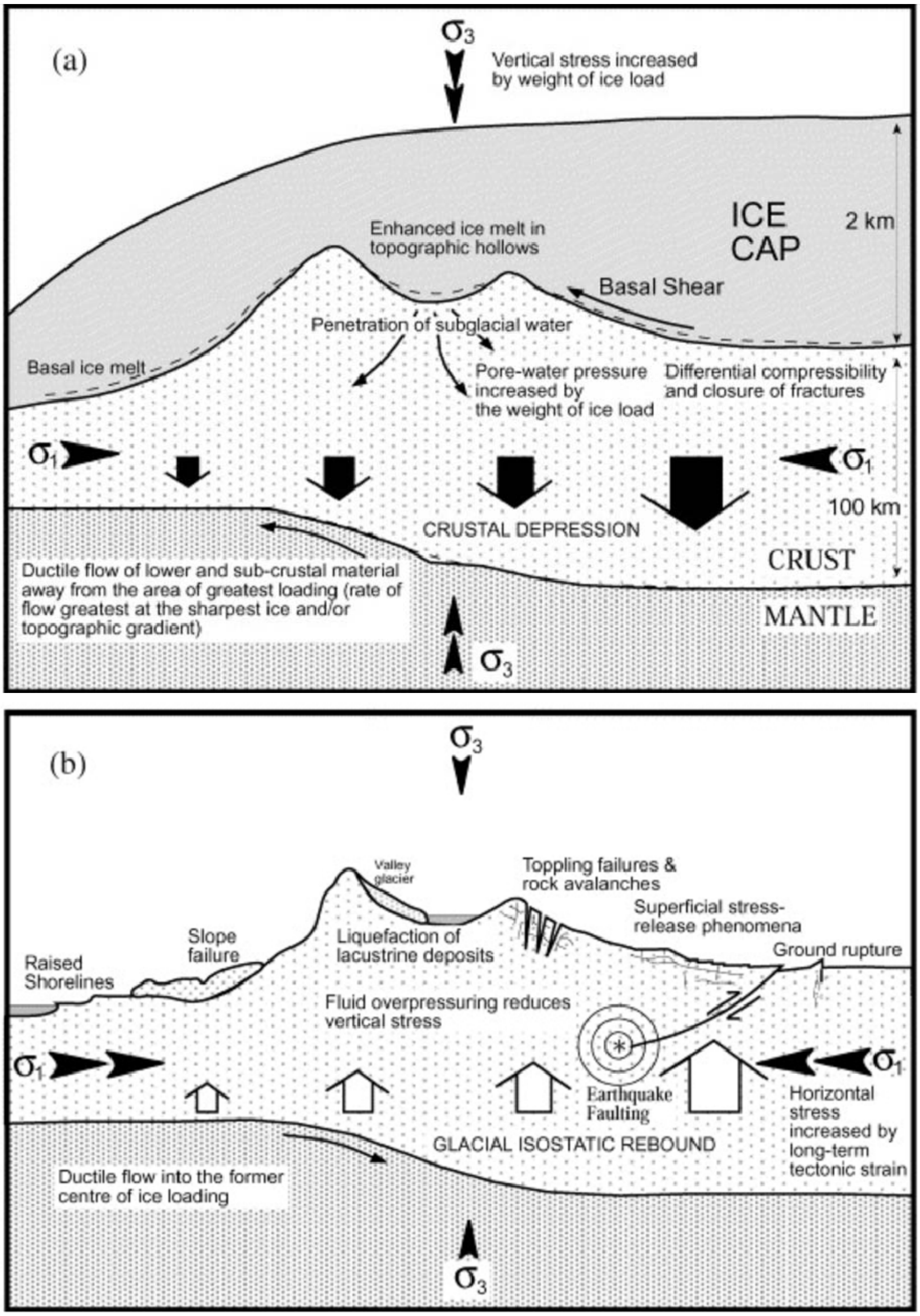


Figure 9. Schematic representation of the impact of glacial loading (a) and unloading (b) on the crust in a region with a compressive stress regime (Reprinted from Stewart, I.S., Sauber, J. and Rose, J., 2000. Glacio-seismotectonics: Ice sheets, crustal deformation and seismicity, Quat. Sci. Rev., 19, pp. 1367-1389 with permission from Elsevier). Note the considerably exaggerated vertical scale difference between the crustal thickness (~100 km) and the ice sheet thickness (~3 km).

Presently, the crust beneath the Antarctic and Greenland ice sheets is depressed in a similar way as previously occurred under the Laurentide and Fennoscandian and Siberian ice sheets. In the former areas, the crust will only be subject to significant rebound if these areas were to be deglaciated in the future (Lund and Näsland, 2009). Modern modelling of glacial rebound (James and Morgan, 1990; Spada et al., 1991; James and Bent, 1994; Wu and Hasegawa, 1996a, b; Wu, 1996, 1997; Johnston et al., 1998; Klemman and Wolf, 1998; Wu, 1999; Wu et al., 1999; Davis et al., 1999, Wieczerkowski et al., 1999; James et al., 2000) illustrates, amongst other things, how the viscoelastic response of the mantle to imposed surface loads can exert significant horizontal stresses. As a consequence, glacial rebound is now widely considered as a potential mechanism for crustal deformation and seismicity not only within limits of former ice sheets but also for several hundred kilometres beyond the limits of glaciations (James and Bent, 1994; Muir-Wood, 2000; Stewart et al., 2000).

Glacial decay and uplift (rebound)

Proposed over a century ago (Jamieson, 1865, 1882; De Geer, 1888) glacio-isostasy is now well documented in previously glaciated regions such as Canada, the US, Fennoscandia, the British Isles and Siberia (refer Bradley et al., 2009). In these areas ancient shorelines are found lying above present day sea-level and the process continues today, around 10-15 kyr after the last deglaciation (e.g. Ekman, 1991; Lund and Näsland, 2009), albeit at a decaying rate (Figure 8). This process is one aspect of GIA and is termed post-glacial rebound (PGR). It is now considered the principal geodynamic process dominating the solid Earth deformation field in these regions (e.g. Johansson et al., 2002; Sella et al., 2007). The speed and amount of post-glacial rebound is determined by two factors: the viscosity or rheology (i.e., the flow) of the mantle, and the ice loading and unloading histories on the surface of Earth. Since mantle viscosity is high, the downwarping and subsequent rebound are slow processes. Post glacial rebound is in fact predicted to continue for another 30 kyr or more in Fennoscandia with a remaining uplift of up to *c.* 100 m before the isostatic effect of the last glaciations is gone (Whitehouse, 2006).

Thus, as glacial periods wane and large ice masses decay and retreat, weight and therefore the vertical stress is removed from the depressed crustal areas relatively abruptly. Initially this leads to elastic rebound, causing crustal uplift, faulting and seismicity. Following this phase of elastic rebound, a longer-term uplift takes place as the displaced mantle flows back over a protracted period into the area beneath the former ice mass, during which the horizontal stresses are more gradually relaxed (Figure 5b and Figure 9). The result is a rising dome of radial crustal extension in the centre of the former ice sheet that expands outward, to 'recover' a forebulge that is collapsing and experiencing land subsidence under crustal contraction as is illustrated by the ongoing lowering in the Baltic states, the Netherlands, southern England and much of the east coast of the USA (Lund and Näslund, 2009). At the same time, there is also a migration of the maximum forebulge towards the formerly glaciated region as the ice sheets retreats. Such models suggest that for a 1 km thick ice load, the rebound is several hundred metres, and the corresponding stresses caused by deglaciation flexure typically reach tens of MPa, which is equivalent to plate-driving stresses, but may be almost an order of magnitude less than stresses imposed by sediment loading on adjacent continental margins (Stein et al., 1989). Within the former ice sheet centre and its forebulge, predicted horizontal strain rates imposed by glacial rebound may be several orders of magnitude greater than observed seismic strain rates (James and Bent, 1994). At greater distances from the load centre, crustal subsidence occurs due to loading of the continental shelves and ocean basins by water returning to the global oceans (Walcott, 1970; Peltier, 1976; Lambeck, 1993a).

The GIA process is seen, not only in the slow rebound of regions of past ice sheets, but also in current global warming trends and the subsequent melting of glaciers producing additional GIA effects worldwide (Lund and Näslund, 2009). For example, rapid glacial rebound is currently occurring in Iceland due to mass loss of the Vatnajökull and other smaller ice caps (Pagli et al., 2007). Since the viscosity of the mantle is a number of orders of magnitude lower beneath Iceland than below *e.g.* Fenoscandia, uplift rates are around 20 mm yr^{-1}, in spite of the much smaller volume of Iceland ice (Lund and Näslund, 2009).

Measurements of PGR/GIA in Britain, Fennoscandia, North America and Canada

Recent uplift and the rates of uplift of glaciated regions can be assessed by both direct and indirect observations. Traditionally, the amount and rate of GIA or PGR has been determined by relative sea-level changes through observations of a number of geomorphological features, these being mainly the elevation and age of raised palaeoshorelines or beaches (e.g. Tushingham and Peltier, 1991; Lambeck, 1993a,b; Davis et al., 1999; Mitrovica, 1996; Lambeck et al., 1998). Inland, above sea-level, lake shorelines and also lake bottoms reveal important data, the effects being similar to those concerning seashores - the bottoms of the lakes gradually tilt away from the direction of the former ice maximum, such that lake shores on the side of the maximum (typically north) recede and the opposite (southern) shores sink and there may be the formation of new rapids and rivers. There have been numerous significant changes to coastlines and landscapes over the last several thousand years as the GIA process has caused the land to move relative to the sea. The effects continue to be significant, with ancient shorelines found to lie above present day sea-level in areas that were once glaciated. However, places in the near-field peripheral bulge outside the former ice margin, which was uplifted during glaciation, the land sinks relative to the sea with ancient beaches found below present day sea-level. This is the case along the east coast of the United States, where ancient beaches are found submerged below present day sea-level and Florida is expected to be submerged in the future (Peltier, 1998). Global Positioning System (GPS) data in North America also confirms that land uplift becomes subsidence outside the former ice margin (Sella et al., 2007). These ‘relative sea-level data’ also demonstrate that glacial isostatic adjustment proceeded at a higher rate near the end of deglaciation than today.

However, the total amount of post-glacial uplift at a site is typically larger than can be inferred from e.g. raised beaches. A large portion of the uplift takes place as the ice sheet starts to decay, prior to the actual deglaciation of a typical site located some distance from the maximum ice margin (Lund and Näslund, 2009). The total maximum amount of glacial rebound that has occurred due to the decay of the Fennoscandian ice sheet is around 800 m (Figure 10a; e.g. Mörner, 1979). This compares with the largest Fennoscandian rebound as inferred from the highest marine limit, which is c. 280 m above sea-level in the Swedish coastal region of the Gulf of Bothnia (Lund and Näslund, 2009).

Another method of making direct observations of on-going post-glacial crustal motions (deformation) of previously glaciated regions has emerged in the use of space geodetic data. Networks of continuously operating, permanent GPS receivers have provided high quality data on post-glacial rebound or GIA in, for example, Canada (e.g. Henton et al., 2006) and North America (e.g. Sella et al., 2007). The present-day uplift motion in northern Europe is also monitored by a GPS network (known as Baseline Inferences from Fennoscandian Rebound Observations, Sealevel and Tectonics (BIFROST) – e.g. Johansson et al., 2002), providing important data for Fennoscandia (e.g. Schernecket al, 1998, 2000; Scherneck et al., 2001; Milne et al., 2001, 2004; Johansson, et al., 2002; Lidberg et al., 2007). A significant advantage of GPS measurements over those of past sea-level changes is that a direct measurement of present-day vertical and horizontal crustal motion is recorded at each site and spatial sampling is not limited to coastal areas. Indirect determination of ongoing GIA has also been undertaken by absolute gravity measurements in Fennoscandia (e.g.; Mäkinen et al., 2005), North America and Canada (Larson and van Dam, 2000; Lambert et al., 2001; Tamisiea et al., 2007); in some cases co-located with GPS measurements. Utilising gravity measurements permits an estimate of remaining uplift in areas where the process is not yet complete (Lund and Näslund, 2009).

Uplift determined by GPS observations for Fennoscandia North America (including Canada) show the same concentric uplift patterns as those derived from sea-level and levelling measurements, with the fastest rebound occurring approximately in the areas where the Laurentide and Fennoscandian ice sheets attained their greatest thickness. The maximum vertical uplift rate measured in this way in the area of the former Laurentide ice sheet is *c*. 13 mm per year (Henton et al., 2006) with GPS data in North America also confirming that land uplift becomes subsidence outside the former ice margin (Sella et al., 2007).

In Fennoscandia, a peak rate of c. 11 mm per year in the north part of the Gulf of Bothnia is recorded, although this uplift rate decreases and eventually becomes negative outside the former ice margin (Scherneck et al. 2001; Johansson, et al., 2002).

In Finland, GIA is measured at a maximum vertical component of 11 mm yr^{-1}, based on GPS measurements. A tectonic component of about 10% of the land uplift (or 1 mm yr^{-1}) is estimated (Ojala et al., 2004). In Fennoscandia, the regional compressive stress field is driven by plate-boundary tectonic processes (Ojala et al., 2004), with stress orientations related to a) the dominant ridge-push/mantle drag related compression and, b) evidence on significant local variations of the surface stress field influenced by the orientation of major fracture zones. However, the level of seismicity following deglaciation strongly suggest that glacial isostatic adjustment is an active process leading faulting in this area, where brittle crust is near the point of failure. Consequently, small changes in the state of stress, like that occurring during deglaciation and glacial rebound (0.1 Mpa) are enough to nucleate earthquakes by the reactivation of optimally oriented pre-existing weaknesses. The post-glacial faults are re-activated old faults and the areas of post-glacial faulting are still the most seismically active areas in Fennoscandia. Further, the association of seismicity with glacial rebound suggests that in areas experiencing diminishing rebound, the level of seismicity decreases over time (Ojala et al., 2004).

As described previously, accompanying vertical crustal motion is a horizontal component of motion during GIA. The BIFROST GPS network has provided evidence that the motion diverges from the centre of rebound, with the largest horizontal velocity found near the former ice margin (Johansson et al., 2002). In Fennoscandia, GPS data record horizontal motions of 2 m yr^{-1} are directed outward from the area of the fastest uplift, with horizontal tectonic motions estimated at less than 1 mm yr^{-1} (Ojala et al., 2004).

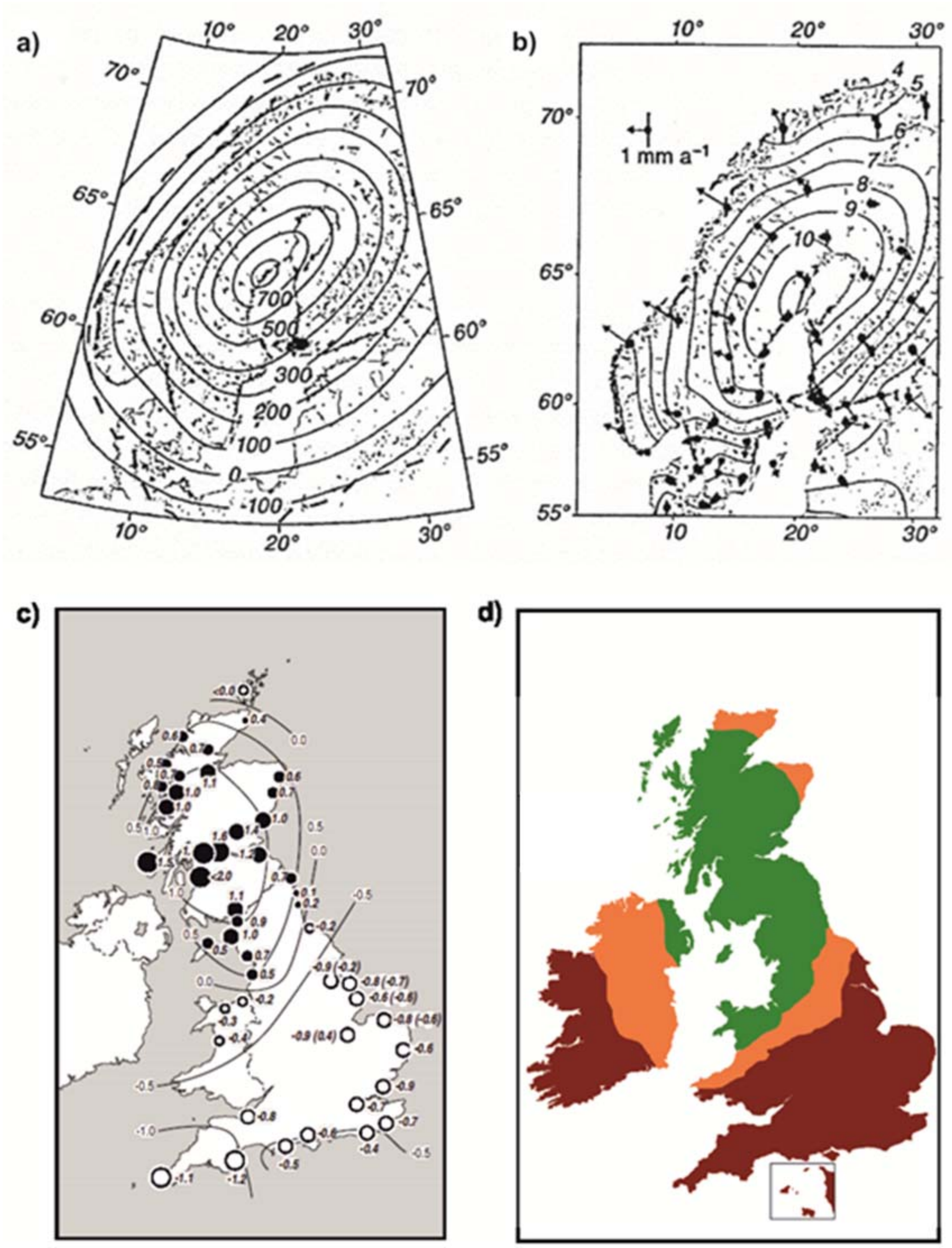


Figure 10. Estimates of glacial rebound and present day crustal deformation in Fennoscandia and Great Britain. (a) The total glacial rebound due to the decay of the Fennoscandian ice sheet. Bold dashed line shows the maximum extent of the ice sheet. Contours are at 100 m intervals. Note that the total amount of rebound is more than twice that inferred from raised beaches formed at the highest post-glacial sea-level (from Lund and Näsland, 2009 and modified from Fredén, 1994 and Lidberg et al, 2007). (b) Present-day crustal deformation over Fennoscandia as observed by continuous GPS measurements within the BIFROST project (Scherneck et al., 2001; Lund and Näslund, 2009). The rebound displays a concentric pattern with a maximum rebound rate of *c*. 11 mm yr^{-1}, located approximately in the centre of the former maximum ice sheet thickness. In the periglacial upwarping process, a horizontal

component of crustal deformation (arrows) is also present, directed outward from the area of maximum depression. Vertical bars represent the error on vertical rebound rate. Horizontal errors in displacement are uniformly small (< 1 mm yr^{-1}). (c) and (d) Late Holocene relative land-/sea-level changes (mm yr^{-1}) in Great Britain, positive values indicate relative land uplift or sea-level fall, negative values are relative land subsidence or sea-level rise. Figures in parentheses are the trends that take into account modelled changes in tidal range during the Holocene (after and based upon Shennan and Horton, 2002). (d) Green shows land which is rising as a result of post-glacial rebound, orange shows stable or near stable conditions and brown shows land which is sinking.

Although small in global terms the ice sheet that covered much of the British Isles at the Last Glacial Maximum (LGM) was large enough for GIA processes to produce vastly contrasting relative sea-level (RSL) changes at different locations (Shennan and Horton, 2002). The size of the British Isles LGM ice sheet means that the glacial isostatic component of RSL change is highly sensitive to shallow earth structure, especially lithospheric thickness and the viscosity of the upper mantle (e.g. Lambeck, 1995; Shennan et al., 2000, Peltier et al., 2002). This is in contrast to, and independent from, rheological constraints derived from regions beneath the more extensive Fennoscandia and Laurentide ice sheets, where RSL changes show more sensitivity to deeper earth structure (e.g. Peltier, 1998), modelling of which contributes to a significantly more complex analysis of relative land- and sea-level movements.

A map of current rates of land- and sea-level change (Figure 10c and d) reveals the highest relative uplift of up to ca. 1.6 mm yr^{-1} (10 cm per century) centred around the area of central and western/north-western Scotland with maximum ice at LGM (Ballantyne et al., 1998; Shennan et al., 2002). Relative land subsidence of up to 5 cm per century occurs in the south and east of England with a maximum in southwest England of ca. 1.2 mm yr^{-1} (Shennan and Horton, 2002). The fulcrum of this re-adjustment is (very approximately) along an imaginary line drawn from just north of Tees Bay in the east to the Dee Estuary in the west (Figure 10c and d). The data suggest that the rate of uplifting of northern England is beginning to slow demonstrably. This means that the eustatic components of global sea-level rise will start to become more pronounced in the region.

The subsidence rates are less than those in an earlier analysis (Shennan, 1989) due to the availability of data permitting a minimum estimate of sediment consolidation (autocompaction). The average effect of allowing for consolidation in this way is equivalent to 0.22 mm yr^{-1}, but there are greater effects in areas with thick sequences of Holocene sediments in the large estuaries in south and east England and where land drainage increases input, increasing the subsidence to between 0.5 and 1.1 mm yr^{-1} (Shennan and Horton, 2002). The effect is most significant for large coastal lowlands, the Fenland and Humber (ca. 0.5 and 0.6 mm yr^{-1}), that were tidal embayments during the mid- to late Holocene.

Comparisons of estimates of crustal velocities within Great Britain based on continuous global positioning system (CGPS) measurements to predictions from a model of GIA show that observed and predicted values for vertical motion show good correlation. The predicted uplift pattern is that of an ellipse, oriented approximately south-southwest to north-northeast, covering most of Scotland and a part of Ireland. The centre of uplift is located in central western Scotland with a magnitude of ~0.6 mm $yr^{-1.}$ The majority of England is subsiding with rates exceeding ~1 mm yr^{-1} in places. While there is a broad similarity between predictions and observations both here and from relative sea-level studies (above), significant differences are observed, which are attributed in part to the contrasting spatial sampling of the two patterns (Shennan and Horton, 2002). The modelling thus indicates that GIA is the dominant geodynamic process contributing to the vertical field (Shennan and Horton, 2002). In contrast, motion of the Eurasian plate dominates the horizontal motion component. These authors adopted a model of plate motion to remove this signal in order to estimate intraplate horizontal motion associated with GIA, but found no evidence of a coherent pattern of horizontal motion in the resulting velocity field.

The direct effect of GIA, both during the glacial depression and post glacial rebound phases, on a GDF will be small. There will be minor adjustments on joints to accommodate the changing local stress fields which may close during loading and relax during rebound. In turn this may influence groundwater flow paths and rates depending on whether joints have reduced or increased in aperture. Modelling would be required to determine whether these changes are significant. Plastic clays are likely to accommodate the changes in stress by deformation rather than joint movement and this is unlikely to have any effect on the isolating characteristics of these rock types.

2.3.3 Neotectonics: GIA and post-glacial seismicity – ‘glacially induced faulting’

A fundamental question in the study of intraplate earthquakes in Laurentia and Fennoscandia is the relative importance of plate tectonics and post-glacial rebound (including both glacial loading and unloading) in earthquake generation. Geological and geophysical evidence exists to support either post-glacial rebound or, tectonic stress, as the dominant cause of these intraplate earthquakes. In order to mitigate more effectively the hazards associated with these earthquakes and to plan for safe storage of radioactive waste in a GDF, it is vital to understand the spatio-temporal variation of the state of stress, the fault potential and the cause of such earthquakes.

Generally, large destructive earthquakes are viewed as occurring along plate boundaries. However, large intraplate earthquakes occur in eastern Canada (up to M_W7) and in Fennoscandia, northern Europe (up to M5), where a significant end-glacial earthquake burst was short lived, probably much less than a thousand years, and confined to the period of the complete melting of the Fennoscandian ice cap (Lagerbäck and Sundh, 2008). Following this flurry of ‘glacial seismotectonics’, flexural stresses that had built up beneath the Fennoscandian glacial load have now diminished to comparable or lower magnitudes with respect to the ambient tectonic stress field, as the depressed lithosphere has gradually relaxed (Johnston et al., 1998; Wu et al., 1999).

In Fennoscandia, during the last stages of the Weichselian glaciations (9,000-15,000 years BP), reduced ice loads and glacially affected stress fields resulted in active faulting with fault scarps up to 160 km long and 30 m high (Kukkonen et al., 2010). Significant post-glacial displacements on basement faults and the seismic disturbance of Late Quaternary sediment sequences have also been documented in Scotland suggesting that during post-glacial times these ancient faults were reactivated by large-magnitude (M_W 6-7) earthquakes (Davenport and Ringrose, 1985; Davenport et al., 1989; Ringrose et al., 1991; Fenton, 1992) and that contemporary seismicity in the Scottish Highlands is concentrated in and around the glacio-isostatic uplift centre, and it is said to be clustered in areas of Late Quaternary shoreline dislocations, faults and palaeoseismic phenomena (Davenport et al., 1989; Musson, 1996). All these settings for intraplate earthquakes and seismicity are far away from present-day plate boundaries. Post-glacial faulting, therefore, indicates that the glacio-isostatic compensation is not only a gradual viscoelastic phenomenon, but also includes unexpected violent earthquakes, often larger than other known earthquakes in stable continental regions (Kukkonen et al., 2010).

Early workers recognised that the upper-crustal response to ice-mass fluctuations was likely to have involved intense seismic activity (de Geer, 1940 cited in Mörner, 1979). However, driven largely by radioactive waste disposal and seismic-hazard concerns in the previously glaciated intraplate shields of North America and Fennoscandia, the role and significance of seismicity in this process and its potential importance as a 'glacial' process in Quaternary science has only really been pursued and recognised in the last few decades (e.g. Mörner, 1978, 1996; Adams, 1989b; Hunt and Malin, 1998; Talbot, 1999; Stewart et al., 2000; Pascal et al., 2010; Hampel et al., 2010). An important intraplate earthquake was the 1811 magnitude 8 New Madrid earthquake that occurred in mid-continental USA and which has been linked to deglaciation (Grollimund et al., 2001).

It is seen from the preceding sections that the last two decades have led to significant growth in our understanding of the past and continuing effects of ice sheets and glaciers on the contemporary crustal deformation and importantly, seismicity. Deformation of the solid Earth is transient and occurs on different spatial and temporal scales. It is dependent upon the rheology of the crust and lithospheric mantle and may arise, for example, from volume fluctuations of lakes, glaciers or ice sheets (e.g. Nakiboglu and Lambeck, 1982; Bills et al., 1994; Peltier et al., 2002; Sauber and Molnia, 2004), from the filling of water reservoirs (Kaufmann and Amelung, 2000) or from earthquakes and the subsequent post-seismic relaxation (Savage and Svarc, 1997; Peltzer et al., 1998; Pollitz et al., 2001, 2008; Hetland and Hager, 2003; Nishimura and Thatcher, 2003; Freed and Bürgmann, 2004; Gourmelen and Amelung, 2005). Since the mantle and the lithosphere continuously respond to the changing ice and water loads, the state of stress at any location continuously changes in time.

The terms 'glacially induced faulting' or 'glacial induced fault' (GIF) used here are generic terms referring to all faulting related to the emplacement or removal of large quantities of ice in glaciers, ice caps or continental ice sheets (e.g. Lund and Näsland, 2009). There is no link to the timing of faulting relative to the evolution of the ice mass. Other frequently used terms include glacio-isostatic, endglacial or post-glacial faults - the latter two being subsets of the definition here and may in some papers be applied to all glacially associated faults despite some being demonstrably associated with glacial advance (refer Munier and Fenton, 2004).

A review of the understanding and investigations of GIFs in North America, Fennoscandia, Russia and Britain (Munier and Fenton, 2004) illustrated how they have almost exclusively been recorded in regions of low to moderate seismicity, i.e. passive continental margin, failed rift or intraplate/craton environments. They generally also only occur in regions where there is no evidence of surface rupture during historical time, or where historical records of seismicity approach the magnitude thresholds for generating surface faulting (Lund and Näsland, 2009). To date all examples of post-glacial faulting have involved reactivation of existing faults and fractures and all large (km-scale) faults currently accepted as being glacially induced are located in Fennoscandia and are almost exclusively reverse faults (Kujansuu, 1964; Lagerbäck, 1979; Olesen, 1988; Munier and Fenton, 2004). The Pärvie Fault is the longest at *c.* 160 km long and with a maximum vertical displacement of 10-15 m (Lund and Näsland, 2009). The GIFs are interpreted to have ruptured as one step earthquakes through a series of old zones of weakness (shear zones), reaching M_W 7-8, around the time the ice was retreating from the respective areas (e.g. Lagerbäck, 1979; Olesen, 1988).

The mechanics of GIF is not well understood because of uncertainties in fault geometries, the tectonic stress state, crustal pore pressures and ice histories (Lund and Näsland, 2009). Two mechanisms are thought possible, with contributions from both likely:

- The vertical stress induced by the ice load decreases the differential stress in the crust, increasing the stability of faults. During glaciations, tectonic strain builds up in the crust under the ice and is later released during deglaciation, when the vertical load is removed. This 'sudden' (in geological terms) release of stored stress is inferred to trigger end-glacial faulting (e.g. Johnston, 1987);

- The horizontal stresses (σh) do not follow the increase in vertical stress (σv), so that as the lithosphere warps, compressional and tensional horizontal flexural stresses are induced which, at some locations are significantly larger in magnitude than the actual ice load (e.g. Johnston et al., 1998). Near and under the ice sheet, σh will be larger than σv all through the glacial cycle, particularly so towards the end of deglaciation (Figure 11). The isostatic rebound of the depressed elastic lithosphere is a much slower process than the retreat of the ice sheet and the corresponding removal of the ice load, leaving remnant high horizontal stresses in the lithosphere, once σv induced by the ice sheet is gone.

The shape of the depression and thus the magnitude and distribution of induced stresses depend on the ice sheet configuration, e.g. spatial extent, surface slope and ice thickness. In the absence of the stabilising effects of the ice sheet, the high horizontal stresses will lead to fault instability and trigger end-glacial earthquakes (e.g. Wu and Hasegawa, 1996b). The glacial-induced shear stresses are not large enough on their own at seismogenic depths to induce faulting. An additional source of deviatoric stress, such as tectonic stress, is required, with the tectonic stress state determining which faults will be reactivated, how and when they will slip and where, in relation to the ice sheet, instabilities will occur (Lund and Näsland, 2009).

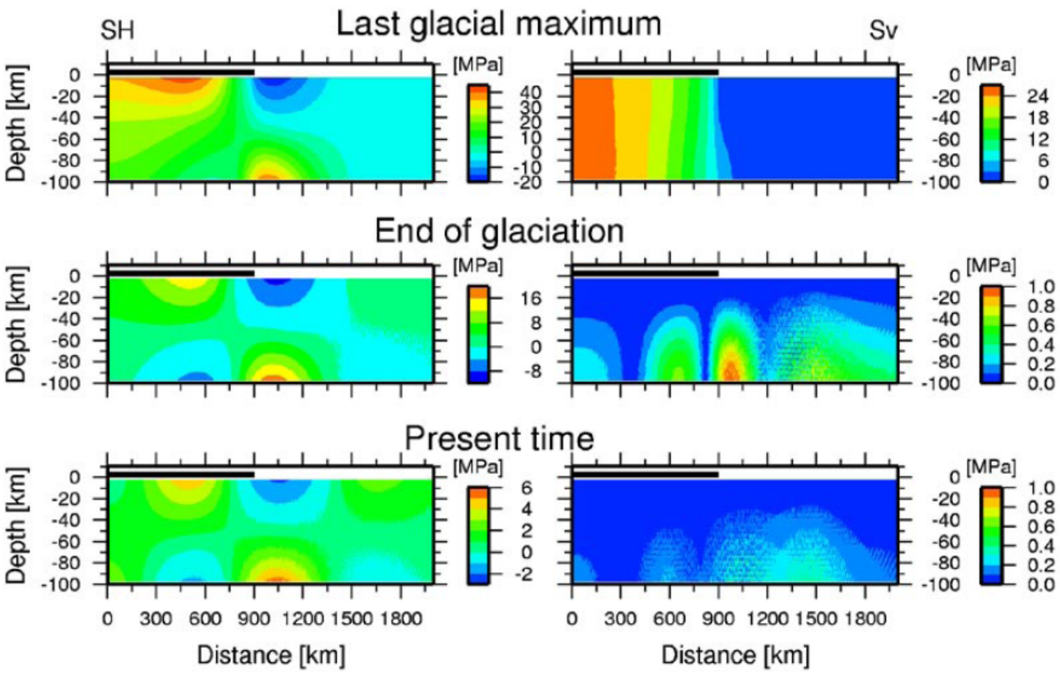


Figure 11. Maximum horizontal (left) and vertical (right) incremental rebound stresses at three different times: LGM, end of glaciation and present (after Lund, 2005; Lund and Näsland, 2009). The extent of the ice load at LGM is indicated by a black bar on top of the plots and the earth model is a simple 2D model with a 100 km elastic lithosphere subject to a generic elliptical cross section ice load with 900 km lateral extent and 25 MPa central pressure at the last glacial maximum.

Pore water pressure in bedrock fractures and pore space is an important component in all faulting related processes. Increased pore pressure decreases the normal stress on a fault and thus decreases the shear stress needed to induce movement on the fault. The ice sheet is expected to increase pore pressure in the crust below, through both the increase in mean stress by the load itself, and through high water pressures at the base of ice body. Increased pore pressures will tend to lower fault stability all through the glacial cycle, although the effect will be greatest at the end of glaciations when the fault stability is generally decreasing (Lund and Näsland, 2009).

2.3.3.1 STATE OF STRESS AND INTRAPLATE EARTHQUAKES

Glacially induced faulting correlates strongly with the growth or decay of ice sheets, but is not, however, limited in time to the actual end of a glacial cycle but may occur at any rapid increase or decrease in ocean volume and/or areal extent during the glacial cycle. The evolution of the ice sheet therefore determines the likelihood of glacially induced faulting at any specific period in time (Lund and Näsland, 2009).

Johnston (1987, 1989) first noted an apparent tendency for large continental ice sheets to suppress tectonic activity because in regions where the maximum tectonic stress is horizontal (that is, in strike-slip and compressive tectonic regimes, but not in tensional regimes), the existence of an ice sheet will tend to increase vertical stresses far more than horizontal stresses, thereby bringing vertical and horizontal stresses more into balance (Figure 9a). The resulting reduced ‘deviatoric stress' promotes the conditions for fault stability, and therefore, inhibits seismic-strain release. He argued that glacial loading could explain the relative aseismicity of the crust beneath the present-day ice caps of Antarctica and Greenland. Numerical models also suggest that fault instability is sensitive to the lateral dimensions of the load with respect to the effective elastic thickness of the plate, providing an explanation for the apparent lack of ‘end-glacial’ surface ruptures in northern America (Johnston et al. 1998).

It was subsequently proposed that on-going tectonic motions will tend to progressively reduce crustal stability the longer the ice sheet is in existence. Where tectonic loading was fast enough in relation to the duration of the ice sheet (as in plate boundary regions), then the crust below the ice sheet would eventually revert to its pre-glacial tectonic state, and seismicity would resume at levels comparable to those that existed before ice sheet growth (Muir-Wood, 1989b and 2000).

The Mohr-Coulomb theory of rock failure suggests that large glacial loads will in general suppress earthquakes, whilst rapid deglaciation would favour earthquake generation. In a review and modelling of slip rate variations on faults during glacial loading and post-glacial unloading and the implications for the viscosity structure of the lithosphere, Hampel et al. (2010) discuss how processes such as post-seismic relaxation or mass fluctuations on the Earth’s surface not only cause deformation that can be measured as continuous displacement field, but also affect the slip behaviour of faults by promoting or delaying earthquakes (e.g. Johnston, 1987; Stewart et al., 2000; Amelung and Bell, 2003; Sauber and Molnia, 2004; Hetzel and Hampel, 2005). Geodetic and seismological data from southern Alaska, for example, indicate that glacier mass fluctuations between 1995 and 2000 were associated both with vertical surface displacements at rates of up to 1 mm yr^{-1} and with changes in the frequency of small to moderate earthquakes (Sauber et al., 2000; Sauber and Molnia, 2004). Field evidence from formerly glaciated regions such as Scandinavia, the northeastern Basin-and-Range Province and the European Alps shows a strong temporal correlation between the melting of the ice-caps and an increase in seismicity at the end of the last glacial period (Lundquist and Lagerbäck, 1976=; Lagerbäck, 1979, 1992; Mörner, 1978, 2005; Olesen, 1988; Byrd et al., 1994; Arvidsson, 1996; Dehls et al., 2000; Ruleman and Lageson, 2002;

Becker et al., 2005; Ustaszewski et al., 2008). A causal relationship between deglaciation and the increased seismicity has been revealed by numerical models that evaluated the response of faults to glacial loading and unloading (Hetzel and Hampel, 2005; Hampel and Hetzel, 2006; Hampel et al., 2007; Turpeinen et al., 2008). Based on these models, the post-glacial increase in fault slip rates can be explained by changes in the differential stress, which describes the difference between the maximum principal stress (σ1) and the minimum principal stress (σ3). During deglaciation and rebound of the lithosphere, the differential stress σ1- σ3 increases, with the consequence that fault slip accelerates.

Post-glacial increase in seismicity has been reported for both thrust (e.g. Lundquist and Lagerbäck, 1976; Lagerbäck, 1992; Mörner, 1978) and normal faults (Byrd et al., 1994; Ruleman and Lageson, 2002). As contractional and extensional tectonic settings are characterized by different orientations of the maximum and minimum principal stresses (σ1-σ3), it may be expected that they respond differently to glacial loading and post-glacial unloading. In contractional tectonic regimes, deglaciation should promote faulting, because the vertically oriented minimum principal stress (σ3) is reduced by the removal of the ice mass. In contrast, faulting should be suppressed in extensional settings because post-glacial unloading decreases the vertically oriented σ1. As numerical models show, however, both types of faults experience a slip rate increase because post-glacial unloading is accompanied by rebound of the lithosphere, which affects the horizontal principal stress in such a way that the net effect is an increase in the differential stress (Hetzel and Hampel, 2005; Hampel et al., 2007; Turpeinen et al., 2008). The major difference between the previous numerical models, which were designed to reflect the tectonic setting of Scandinavia and of the Basin-and-Range Province, was the rheological layering of the lithosphere: the thrust fault model (Turpeinen et al., 2008) was characterized by a weak lower crust overlying a strong lithospheric mantle, because such a viscosity structure has been inferred for cratons such as Scandinavia and mountain belts such as the Himalayas based on laboratory experiments, geophysical data and numerical modelling (e.g. Brace and Kohlstedt, 1980; Chen and Molnar, 1983; Beaumont et al., 2004; Burov and Watts, 2006; Klemperer, 2006). In contrast, the normal fault models (Hetzel and Hampel, 2005; Hampel et al., 2007) reflected the viscosity structure of the lithosphere beneath the actively extending Basin-and-Range Province, where the lithospheric mantle has a lower viscosity than the lower crust (e.g. Kaufmann and Amelung, 2000; Amelung and Bell, 2003; Nishimura and Thatcher, 2003; Gourmelen and Amelung, 2005).

In seismotectonic terms, the deglaciated upper-crustal region should be in tension (normal faulting) and the unglaciated forebulge should be in compression (thrust faulting), though compressive stress regimes often prevail within formerly glaciated areas due to on-going regional tectonic compression. Nevertheless, across the margins of former ice sheets there are often marked changes in faulting style. For example, the focal mechanisms of the larger events indicate that thrust faulting is the dominant mechanism in Eastern Canada. However, along the northeast coast of Baffin Island, focal mechanisms are of the normal-fault type while in north-eastern United States, strike-slip faulting predominates (Stein et al., 1989; Hasegawa and Basham, 1989).

In general, the short-term response to ice fluctuations is similar to the Earth's response to fluctuations in water reservoirs, with the subsequent increase or decrease in seismicity dependent on the pre-existing stress state (Stewart et al., 2000). When the ice fluctuations occur on the spatial scale, and magnitude, of the Late Pleistocene glaciation and deglaciation, however, the viscoelastic response of the Earth (especially the mantle) causes significant changes in crustal deformation and earthquake activity that is spatially extensive and temporally complex. The regions of greatest ice thickness and the regions marginal to the

Late Pleistocene ice sheets indicate the most dramatic evidence of earthquake faulting. The mantle response to these large ice mass fluctuations and the change in mass between the oceans and land caused, and continues to cause, measurable crustal deformation at hundreds of kilometres from the ice margins.

2.3.3.2 Post-glacial tectonics of the Scottish glacio-isostatic uplift centre

A detailed re-evaluation of post-glacial fault movements, seismic activity, possible earthquake triggered soft sediment deformation (seismites) in Late Quaternary soft sediments (glaciofluvial outwash sands and gravels, glaciolacustrine deposits, fluvial/lacustrine sediments) and shoreline sequences suggests that the period of deglaciation and the early Holocene was more seismically active than the mid to Late Holocene (Firth and Stewart, 2000). Post-glacial shoreline dislocations around the periphery of the Scottish glacio-isostatic uplift centre imply metre-scale vertical crustal movements, probably on basement faults. Within the uplift centre itself, geological studies have reported extensive post-glacial reactivation of short (1-14 km long) fragments of major basement faults (Davenport and Ringrose, 1985; Davenport et al., 1989; Ringrose et al., 1991; Fenton, 1992). Although many of these faults exhibit metre-high scarps that are consistent with the scale of vertical offsets inferred from shoreline dislocations, many of the most prominent structures are considered to display large (10^1-10^2 m) horizontal post-glacial offsets (refer Firth and Stewart, 2000). The very significant post-glacial offsets attributed to the fresh-looking faults in the Scottish Highlands is taken as an important indicator that they represent seismotectonic, rather than glacio-isostatic, structures, akin to the large early Holocene earthquake faults of the Lapland Fault Province, northern Fennoscandia (Muir-Wood, 1989b).

Post-glacial strike-slip faults have been inferred from the former glaciated shield of Fennoscandia (Talbot et al., 1989), but appreciable slip has not been documented from such structures. Instead, post-glacial faults in the Lapland region are predominantly thrust faults accommodating vertical displacements ranging in scale from a few metres up to 30 m (Muir-Wood, 1989b, 1993; Dehls et al., 2000). Shorter (1-10 km long) normal faults with metre scale (<5 m) vertical displacements are reported from uplands covered by smaller ice caps, such as Ireland (Mohr, 1986; Knight, 1999), the French and Swiss Alps (Geiger et al., 1986; Sébrier et al., 1997), the Cantabrian Mountains, Spain (Alonso and Corte, 1992), and the Peruvian Andes (Ego et al., 1996). Evidence suggests therefore that dip-slip post-glacial faults accommodating vertical crustal movements predominate in former glaciated domains, and in this respect, strike-slip post-glacial faulting of the western Scottish Highlands would appear to represent an anomalous tectonic environment (Callum and Stewart, 2000).

Although sea-level studies indicate that glacio-isostatic uplift is probably still continuing (Shennan, 1989; Shennan and Woodworth, 1992), it remains uncertain if this rebound is significant enough to modulate or even affect the crustal stress regime. Available data from the Scottish Highlands was consistent with the prevailing view that deglaciation of the region was accompanied by enhanced seismotectonic activity. Firth and Stewart (2000) argued that in Scotland, unlike Fennoscandia, it is not possible to demonstrate this association unequivocally, proposing that the large-scale lateral displacements formerly proposed cannot be justified. Rather, all post-glacial fault movements appear to be limited to metre-scale vertical movements along pre-existing fault lines. In addition, it is argued that the Younger Dryas ice advance may have produced localised crustal redepression but not the more widespread impact formerly proposed. Both tectonic and post-glacial rebound stresses, however, may be needed to explain the contemporary seismotectonics of the Scottish Highlands.

2.4 TECTONIC RELATED UPLIFT AND SUBSIDENCE: POTENTIAL IMPACTS ON A GDF OVER THE NEXT ONE MILLION YEARS

Conventional analyses of passive margin stratigraphy commonly assume that eustacy (changes in sea-level) represents the primary control on sedimentary architecture (Stoker et al., 2005b). It has been shown that in the relatively starved, deep-water, Neogene basins along the north-west European margin, eustacy alone cannot account for the stratal patterns and deep-water unconformities observed in the shelf-margin and basinal sediments (Shannon et al., 2005; Praeg et al., 2005; Torbjørn Dahlgren et al., 2005), nor the timing of change. Instead, the development of the stratigraphy primarily reflects a regional mid- to late-Cenozoic tectonic control that has driven changes in sedimentary and oceanographic dynamics on the north-west European margins and Atlantic region (Stoker et al., 2005a; Praeg et al., 2005). The cumulative effects of tectonism on sediment supply and deep-ocean circulation have latterly been joined by those of Quaternary climate change (Stoker et al., 2005b).

Until recent times, the Cenozoic post-rift development of the north-east Atlantic margins has generally been classed as passive. However, it has been shown that the configuration and evolution of the north-west European margin, including the onshore and offshore areas of the United Kingdom Continental Shelf (UKCS), have been significantly modified by late Palaeogene and Neogene epeirogenic and compressive movements (e.g. Doré et al., 1999, 2002a; Williams et al., 2005; Holford et al., 2008). These structures demonstrate an evolution that has been anything but passive, subjected to various and repeated forces related to ridge-push and Alpine collisional forces during the Cenozoic (although alternative explanations attribute the formation of the anticlinal structures along the Norwegian margin solely to differential loading by Plio-Pleistocene sediment wedges, e.g. Kjelstad et al., 2003).

Ambiguity concerning the causes of compressional tectonism is, in part, related to uncertainty over the number and age of compressive phases recorded by uplifted regions and anticlinal structures. A compilation of age estimates from a number of studies along the north-west European margin implies a record of contractional deformation that was almost continuous from the late Palaeocene to the Pleistocene (Boldreel and Andersen, 1995; Vågnes et al., 1998; Lundin and Doré, 2002; Ritchie et al., 2003; Johnson et al., 2005). Common to most age estimates is an intensification of contractional deformation during the Miocene, resulting in the initiation of new structures as well as the reactivation of older (Palaeogene) structures.

Epeirogenic movements have resulted in a series of broad domes along the north-west European margin, recording kilometre-scale uplift (Rohrman and van der Beek, 1996; Japsen and Chalmers, 2000), as well as in offshore episodes of rapid differential deepening, also of kilometre-scale (e.g. Cloetingh et al., 1990; Vanneste et al., 1995). The domal uplifts include southern Norway and Britain–Ireland (Rohrman and van der Beek, 1996), each of which provides evidence of two main uplift phases, in the early and late Cenozoic (see Dore et al., 2002). Over a comparable period, compressional movements have resulted in the formation of inversion anticlines on parts of the margin and over the Alpine Foreland region, some of which may date back to the late Palaeocene, but the major movements on which date from Miocene and younger Plio-Pliestocene times (Tate, 1993; Boldreel and Andersen, 1995; Tate et al., 1999b; Lundin and Doré, 2002b; Ritchie et al., 2003; Johnson et al., 2005).These are:

- Two phases of tilting (coeval subsidence and uplift, rotations <1° over distances of 100s of kms), causing the basinward progradation of shelf-slope wedges from elongate uplifts along the inner continental margin and from offshore highs:
    - the early Cenozoic (late Palaeocene to Eocene; c. 60-50 Ma);

   - the late Cenozoic (early Pliocene to present; <4 +/- 0.5 Ma);
- A single phase of sagging, with rotations of up to 4° over distances of <100 km) in the late Palaeogene (late Eocene to early Oligocene; c. 35-25 Ma).

This chapter has considered the processes and consequences of uplift and subsidence. Whilst tectonic events have been on-going more or less continuously throughout the Cenozoic in some area or other of the north-east Atlantic margin, in the UKCS area, there have been quite distinct episodes. Given the Cenozoic spans the periods 65.5 Myr to present, this represents a major epeirogenic uplift episode on average once every 22 million years, which is longer than the 1Myr period under consideration in this study.

The processes occurring in areas undergoing subsidence and deposition of thick wedges of sediments, such as those developed during late Neogene times, are likely to increase the further isolation of the waste GDF. It is the processes of uplift and accompanying erosion that will lead to potential unroofing and eventually exposure of any GDF structure and which require careful assessment. Most of the factors above will not significantly affect a GDF in any rock types in any location in the UK over the next one million years. However a number of issues warrant noting here, which are largely related to the regional stress field and the slow changes to it. These are brought about through a number of factors including:

- Plate tectonics - Even the UK suffers the effects of distant plate tectonics, as the stress field readjusts itself, leading to seismicity on major fault lineaments and formation of new and reactivation of old fractures. Although a lot of plate-tectonic-related seismicity occurs far deeper than GDF depths, it may still be an issue at 200 to 1000m depth, through earthquakes and fracture formation/reactivation.
- Isostatic readjustment, either as a result of tectonics (foreland basins and mountain ranges forming bulges), or similar features caused by ice loading as a result of glacial loading and unloading. This occurs in two ways, as a result of tectonic activity (e.g. through formation of foreland basins and bulges) and as a result of glacial loading and unloading. The main issues associated with the latter include:
   - Changes in the regional stress field which can lead directly to fracture formation and reactivation. Geological deformation structures can occur within the 200 to 1000 m depth of a GDF, and can occur in different ways depending on the geological setting. For instance, fractures in argillaceous rocks may form and almost immediately heal themselves, whereas fracturing in crystalline rocks may lead to open fractures.
   - Changes in the hydrology and hydrogeology can occur at the surface/ice interface where an ice sheet can effectively 'switch off' groundwater recharge to the subsurface, leading to changes at depth in the hydrogeology associated with fractures. Conversely, increased pressure at the base of an ice sheet can eventually lead to melting of the ice, resulting in the 'switching back on' of groundwater recharge, again resulting in changes in the hydrogeology associated with faults and fractures at depth.
   - Mineralisation – fracture formation and fracture evolution is typically accompanied by mineralisation due to mineral-rich fluids flowing through the rock matrix (where it can) and along the fracture planes themselves. In fact, in some geological settings fracturing does not occur without some

form of mineralisation. The extent and type of mineralisation is affected by the hydrogeology and size, distribution and extent of fractures. As described later in the main text, mineralisation can effectively lead to the formation of barriers to fluid flow.

Any of the processes described in this chapter may lead to an increase in relative ground level above sea level. Such changes may lower the base level to which groundwater flows are naturally directed. This may in turn lead to changes in the groundwater flow pattern, including higher heads, higher flow rates, longer and deeper flow paths. These in turn may lead to changes in the rock properties and groundwater chemistry that include dissolution and/or deposition of minerals in fractures and pore spaces, changes in solute concentrations and changes to the redox state. All of the potential changes may have an effect on a GDF if it is sited in a location that is affected by them, however, it should be noted that most of the uplift processes described in this chapter occur over long time scales and that they therefore happen relatively slowly. Conversely any of the process that lead to a decrease in relative ground level above sea level are likely to lead to a rise in base level. This is likely to result in lower heads, lower flow rates, shorter and shallower flow paths.

3 UK Seismicity, Tectonic History and Volcanism

3.1 INTRODUCTION

The UK lies on the northwest European shelf and at the northeast margin of the North Atlantic Ocean. In the context of plate tectonics, the UK lies in the northwest part of the Eurasian plate far from any plate boundaries and is characterised by low levels of seismicity and low seismic hazard. The continental crust of the UK formed over a long period of time and has a complex tectonic history, which has produced much lateral and vertical heterogeneity through multiple episodes of deformation (Woodcock and Strachan, 2000), resulting in widespread faulting. The major tectonic blocks and structural provinces that make up the upper part of the crust under the UK are shown in Figure 12 along with both instrumental seismicity (1970 to present) for earthquakes with a moment magnitude, or M_w, (a new, universally applicable scale for measuring the severity of an earthquake which is approximately equal to M_L and M_S) >2.0 and historical seismicity (pre-1970) for earthquakes with M_w >3.0. There are some apparent variations in the spatial distribution of seismicity throughout the UK. In general earthquakes occur in a north-south band along the length of Britain, mainly along the western flank. This band gets wider moving south. The northeast of Britain, the northwest Atlantic margin and Ireland all show an absence of notable seismicity. The band of earthquakes on the UK mainland cuts through the geological terrane boundaries, most of which run northeast southwest. Onshore activity is quite distinct from the seismic activity in the North Sea rift zone. Historical evidence shows that significant earthquakes can affect the south and east of the UK, although there is little instrumental evidence for such events. In Scotland, a strong correlation between seismicity and the expected area of maximum glacio-isostatic uplift has been noted by a number of authors, including Musson (1996).

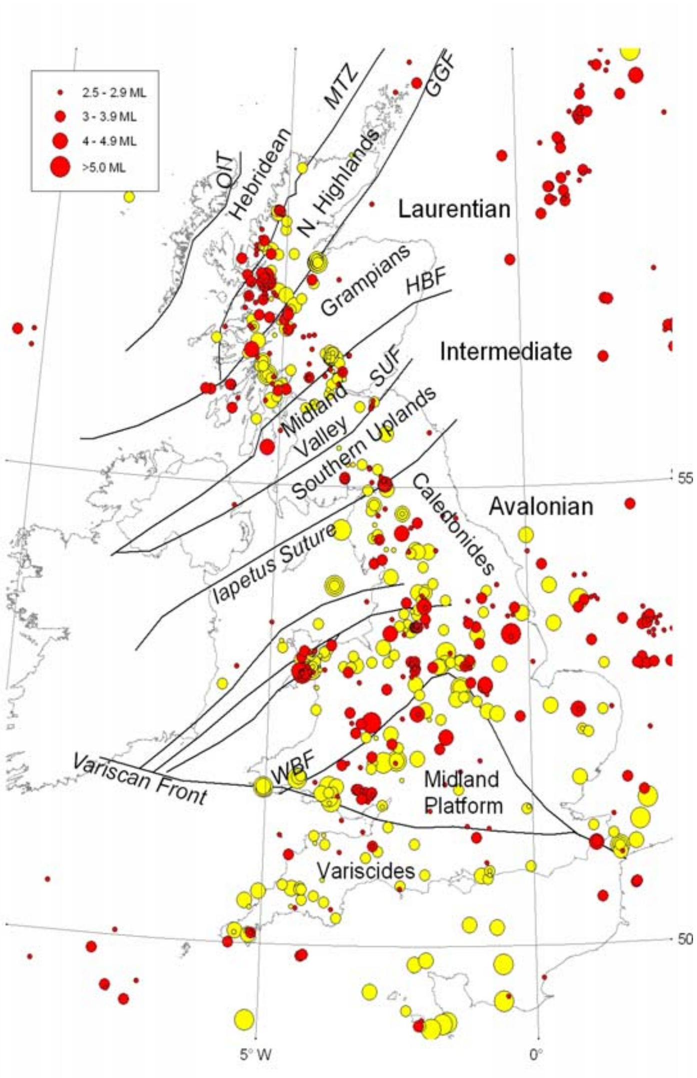


Figure 12: Instrumental (red) and historical (yellow) seismicity of the British Isles from the British Geological Survey earthquake catalogue (Musson, 1996). Earthquake symbols are scaled by magnitude. Geological terranes after Bluck et al. (1992) are also shown. Major faults corresponding to terrane boundaries are abbreviated as follows: Outer Isles Thrust (OIT); Moine Thrust (MTZ); Great Glen Fault (GGF); Highland Boundary Fault (HBF); Southern Uplands Fault (SUF); Welsh Borderland Fault System (WBF).

Reconstructions of plate motions show that during the Phanerozoic the northern part of the British Isles was located at the passive margin of Laurentia, while the southern part was located at the subducting margin of Avalonia. North of the Highland boundary fault the crust is Laurentian, while south of the Iapetus Suture Zone in England and Wales the crust is Avalonian. The closure of the Iapetus Ocean during the Caledonian Orogeny (460-420 Ma) then resulted in the juxtaposition of the two, separated by an intermediate accreted zone in between. Bluck et al. (1992) divides the British Isles into a number of fault-bounded basement blocks or terranes. The amalgamation of these terranes during the Caledonian Orogeny affected the area extending from the Moine Thrust in the northwest to the Welsh Caledonides in the south, resulting in a dominant structural trend that is approximately northeast-southwest. A wedge-shaped basement block of Proterozoic crust called the Midlands Platform dominates much of Southern Britain (Pharaoh et al., 1993), and is terminated by the Variscan Front to the south and Welsh Caledonides to the north. Structures trend northeast in the western part, but northwest in the eastern part of the front. South of the Variscan Front are the strongly deformed Palaeozoic rocks of southern Britain. Structure in the fold belt is generally east/southeast.

Tectonic earthquakes result from sudden slippage along pre-existing faults and are the result of a stick–slip frictional instability on those faults (Brace and Byerlee, 1966). Thus, the earthquake is the ‘slip’ and the ‘stick’ is the interseismic period of elastic strain accumulation. However, no British earthquake recorded either historically or instrumentally has produced a surface rupture and typical fault dimensions for larger UK earthquakes are of the order of 1 to 2 km. It is, therefore, difficult to accurately map earthquakes to specific faults, particularly at depth, where the fault distributions and orientations are unclear, given the large uncertainties involved. A number of studies, for example Ottemoller and Thomas (2007), use the alignment of earthquakes from a specific sequence, along with fault plane solutions, to identify causative faults.

Parameters for historical earthquakes were derived by Musson (1994). Parameters for earthquakes after 1970 were derived from instrumental recordings on the seismograph networks. Completeness of this catalogue varies with time. Musson and Sargeant (2007) estimate that the instrumental catalogue from 1970 to present is complete down to magnitude 3, whereas the magnitude of completeness of the historical catalogue increases with time before present. From 1970-1850 this is 3.5 for all of mainland Britain, whereas prior to 1700 it is at least 4.5.

3.2 DRIVING FORCES FOR PRESENT DAY DEFORMATION

The underlying cause and distribution of earthquake activity in the British Isles is not clearly understood. Main et al. (1999) suggest that the observed neotectonic uplift combined with a direction of maximum (regional) stress deduced from earthquake focal mechanisms supports the theory that deformation is dominated by glacio-isostatic recovery. More recently, Bott and Bott (2004) and Arrowsmith et al. (2005) argue the earthquake activity is a response to an underlying hot, low-density anomaly in the upper mantle that protrudes south from the Iceland plume that provides dynamic support for observed topography despite relatively thin crust. Further evidence for this is provided by Davis et al (2012).

Earthquake source mechanisms provide both fault geometries and principal stress directions that can be used to constrain our understanding of the driving forces of current deformation. Baptie (2010) presents a compilation of focal mechanisms for UK earthquakes in the magnitude range 3 to 5.4. The focal mechanisms (Figure 13) are found to be dominantly strike-slip with northwest-southeast compression and northeast-southwest tension, or reverse, with northwest-southeast compression. In many cases there is also an oblique component to the slip. Pressure (P) and tension (T) axes represent the axes of maximum shortening and maximum extension on a fault plane respectively. In the UK P- and T- axes from individual solutions are relatively well constrained in azimuth, though less so in dip, with P-axes orientation for most events clustering between north and north-west, indicating sub-horizontal compression. However, some spatial variation in P- and T-axes orientation is also apparent, with near north-northeast compression and east-west extension in north-west Scotland, changing to northwest-southeast compression in England and Wales.

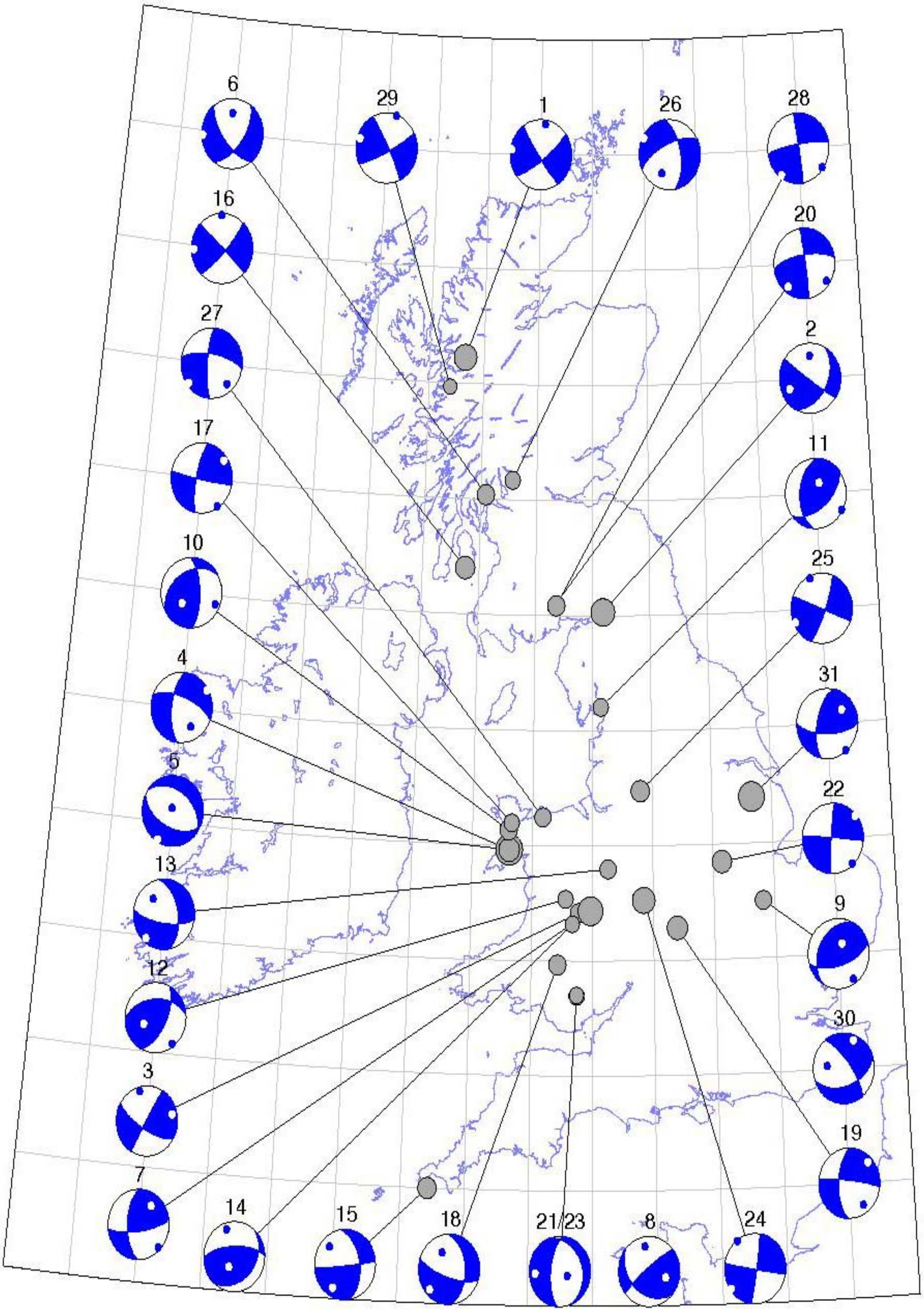


Figure 13: Focal mechanisms for all earthquakes used by Baptie (2010). The compressional quadrants are indicated by the blue shaded areas. P- and T- axes are indicated by the black and white circles respectively.

Figure 14 shows best-fitting stress tensors for two different subsets of the data, suggesting that there is a significant difference in the stress state in northwest Scotland and in England and Wales. Calculated σ_1 directions for England and Wales are northwest-southeast, consistent both with existing stress data and expected stresses from first order plate motions. By contrast, the inversion results for northwest Scotland show near east-west extension with possible σ_1 and σ_2 directions lying in a north-south band, and that the magnitudes of σ_1 and σ_2 are similar. The clear difference in the stress inversion data between northwest Scotland and England and Wales suggests that the principal stress directions expected from first order plate motions have been modified in Scotland by local stress conditions from a second order process.

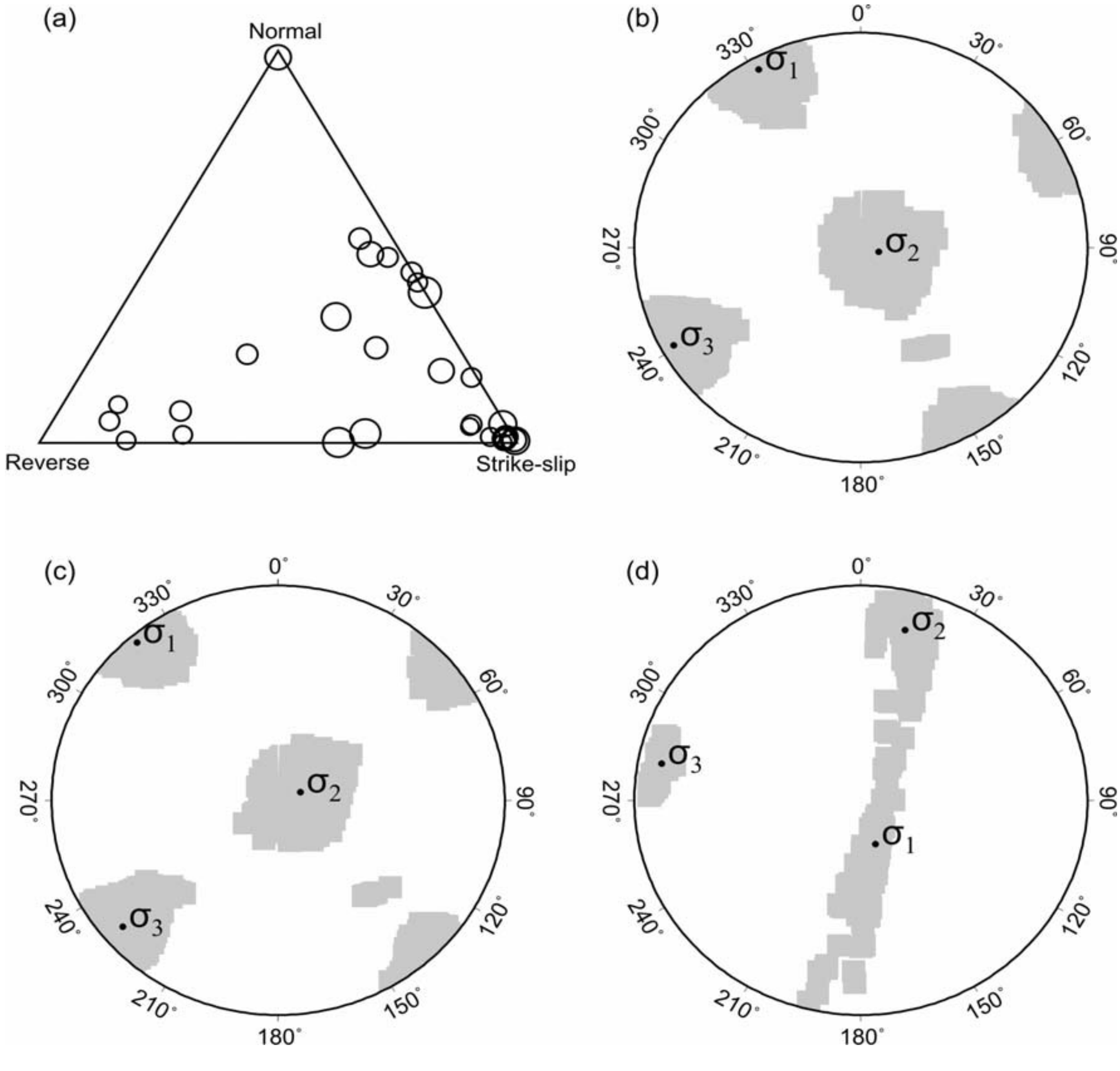


Figure 14: (a) Ternary diagram showing the different components of slip for UK earthquakes. Best fitting stress tensors obtained for: (b) all focal mechanisms; (c) using focal mechanisms for England and Wales only; (d) using focal mechanisms for northwest Scotland only. The 95% confidence intervals are indicated by the shaded areas. From Baptie (2010).

First order intraplate stresses depend mainly on the same forces that drive plate motion. This can result in a uniform stress field over large areas. In the UK, these forces are generated at the Mid-Atlantic ridge due to gravitational effects acting perpendicular to the spreading ridge, and, to a lesser extent, forces resulting from the collision of Africa with Europe. This is expected to result in a prevailing northwest to north-northwest orientation for σ_h, the maximum horizontal compressional stress.

A second source of crustal stress in the UK is vertical uplift from either epeirogeny associated with a hot underlying mantle (as discussed in Chapter 2) or with glacio-isostatic adjustment (GIA). Maximum ice thickness is estimated to be >2 km (Ballantyne et al., 1998; Milne et al., 2006), and there is a good correlation between the spatial extent of the seismicity in northwest Scotland and the region of maximum ice thickness, suggesting that this could be an important factor in the seismotectonics of the UK. Most of our understanding of the rates and patterns of post-glacial uplift in the UK has been determined from long-term estimates of sea-level changes which have been used to constrain quantitative models of isostatic adjustment (Shennan, Bradley et al., 2006b). Uplift rates are around 2 mm yr^{-1} in Northern Britain, which will result in curvature dependent bending stress along the axis of the uplift. Stein et al. (1989) model the effect of a 1 km thick ice sheet and find lithospheric stresses of a few tens of MPa, which is similar to that due to ridge-push. This deglaciation flexure should give rise to tensional stress acting in all directions in the shallow part of the lithosphere under the deglaciated region and compression in the unglaciated region. However, a compressive stress regime may also occur within the deglaciated region due to regional tectonic compression.

3.3 EARTHQUAKE HAZARD IN THE UK

Seismic hazard can be simplified in terms of the spatial distribution of earthquakes or active faults in a given region, the magnitude recurrence relationship for those earthquakes and the decay of earthquake ground motions as a function of distance. These can be combined to give a probabilistic estimate of difference levels of ground motion within a given period of time. The first probabilistic seismic hazard map for the UK was produced by Musson and Winter (1997). The UK was also included in the Global Seismic Hazard Assessment Program (GSHAP, Grunthal et al, 1996) and Seismotectonics and Seismic Hazard Assessment of the Mediterranean Basin (SESAME) (Jiménez et al, 2001). Musson and Sargeant (2007) published seismic hazard maps for the UK for seismic zoning in relation to Eurocode 8 (Figure 15). In general, these studies all show that seismic hazard in the UK is low.

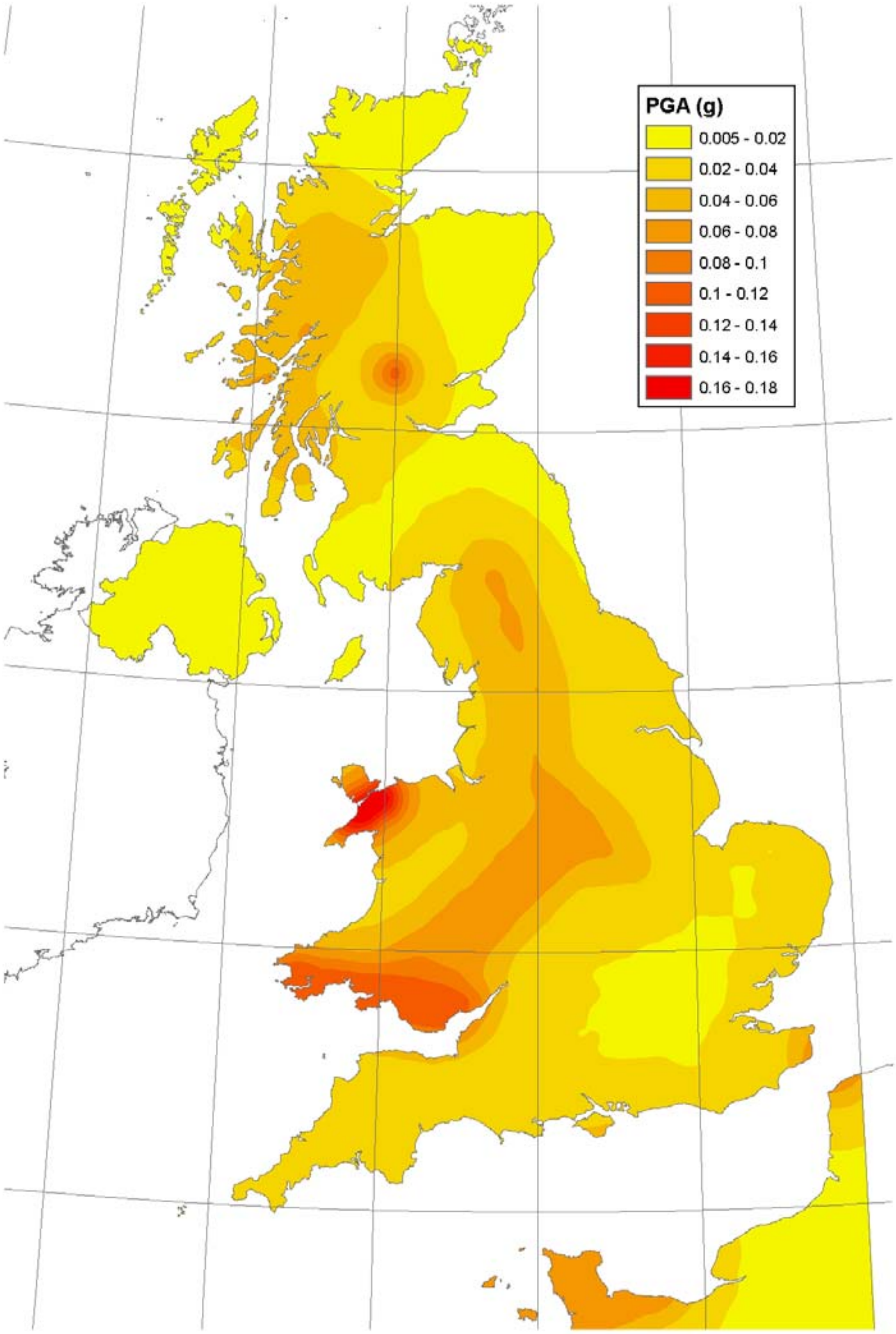


Figure 15: Hazard map showing peak ground accelerations (g) with a 10% probability of being exceeded for a 2,500 year return period (Musson and Sargeant, 2007).

Earthquakes follow a well-established frequency-magnitude relationship (Gutenberg-Richter, 1944), which shows that in a given period of time the number of events of a given magnitude will be ten times less frequent than those of one unit of magnitude less. This means that large earthquakes are less frequent than small ones. The Gutenberg Richter relationship for the British Isles from the instrumental earthquake catalogue (Figure 16) gives an estimate that on average, the UK experiences a magnitude 4 earthquake or greater every 3-4 years and a magnitude 5 earthquake roughly every 15-20 years. As magnitude increases further, recurrence intervals become more uncertain because of sparse observations.

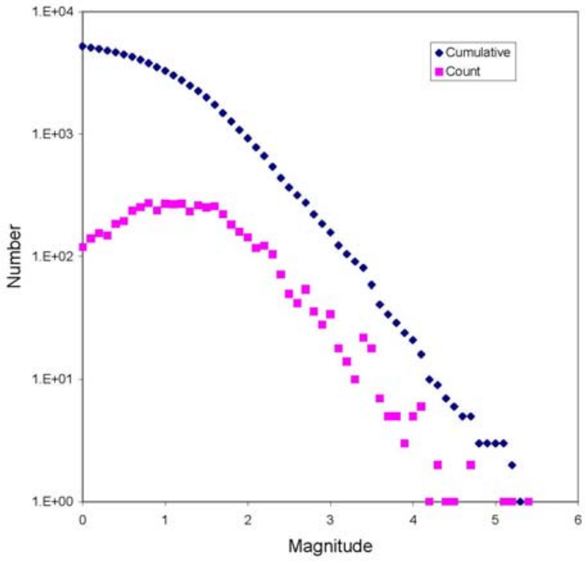


Figure 16: Gutenberg-Richter plot for post-1970 UK seismic showing both the number of events of given size and the cumulative number of that size or greater.

Estimating the largest earthquake that can be expected in the British Isles is difficult because of the low seismicity rates and the limited period of observation. Ambraseys and Jackson (1985) suggest that the maximum possible event may be as low as 5.5 M_w, whereas Main et al (1999) calculate a maximum magnitude of 6.3 M_w. Musson and Sargeant (2007) define a weighted maximum magnitude distribution of between 5.5 - 6.5 M_w, with the highest weighting on magnitude 6 M_w. Earthquake magnitudes scale with fault dimensions and the largest earthquakes (M>8) require fault lengths of hundreds of kilometres. Such faults are usually only found at plate boundaries. As magnitude decreases so do the fault dimensions so earthquakes with magnitudes of around 7 M_w typically have a mean fault length of around 50 km (Wells and Coppersmith, 1994). This places some constraint on maximum magnitude, and although there are a few examples of intraplate earthquakes with magnitudes in excess of 7 M_w (e.g. Hough et al, 2000), maximum magnitudes in the UK are unlikely to exceed 6.5 M_w.

Historically, the largest known British earthquake occurred near the Dogger Bank in 1931, with a magnitude of 5.5 M_w. Although it was 60 miles offshore it was still powerful enough to cause minor damage to buildings on the east coast of England. Probably the most damaging UK earthquake was the Colchester earthquake of 1884 (4.6 M_w). Some 1200 buildings needed repairs and, in the worst cases, walls, chimneys and roofs collapsed.

The largest instrumentally recorded earthquakes occurred on the Lleyn Peninsula in North Wales (Turbitt et al, 1985), with a magnitude 5 M_w in 1984, and in Market Rasen, Lincolnshire (Ottemöller and Sargeant, 2010) with a magnitude 4.5 M_w in 2008. They were both felt throughout England and Wales and into Ireland and Scotland, causing widespread public alarm and some minor superficial damage close to the epicentres. On 28 April 2007, a magnitude 4.3 M_w earthquake near Folkestone (Sargeant et al, 2008) resulted in emergency measures being taken by local authorities, power outages, transport disruptions and localised damage of a severity not seen in the UK in at least fifty years. Given the proximity of this event to other larger earthquakes (of the order of magnitude 5.5-6) (Varley, 1996) in the Dover Straits region in historical times (1382 and 1580) the probability of a future event that causes even more disruption over a wider area is not insignificant. This is also true for densely populated parts of the Midlands and Wales. The relevance of this understanding will be discussed in section 3.5.

3.4 VOLCANISM

The UK is not a volcanically active area, the last volcanic episode being that related to the opening of the Atlantic Ocean and which formed the basalt lava flows of Antrim and the volcanic centres of the Hebrides. No volcanic activity has occurred in the UK in last 60 My (e.g. Pearson et al 1996). Given the UK's stable, mid plate location it is highly unlikely that volcanism will occur during the next several 10's of million years and it is considered improbable that any volcanism will occur in the UK over the next one million.

3.5 SEISMICITY AND TECTONICS: POTENTIAL IMPACTS ON A GDF OVER THE NEXT ONE MILLION YEARS

Earthquake activity presents a number of hazards for radioactive waste disposal: rupture displacement hazard; vibratory hazard (ground shaking); and secondary hazards (e.g. changes to groundwater). Fault displacement hazard refers to the danger of physical movement along a fault plane disrupting the waste emplacement; one can distinguish between principal fault hazard (movement along the fault plane of the earthquake) and distributed fault hazard (secondary movement at some distance from the principal fault plane) (Stepp et al., 2001). Vibratory hazard concerns the possibility of damage due to strong shaking. Secondary hazard due to seismic disruption of groundwater patterns is also possible.

3.5.1 Rupture Hazard

The rupture hazard for a shallow GDF in a low seismicity intraplate region such as the British Isles is very low. Larger earthquakes tend to nucleate at mid-crustal depths, where the strength of the rocks is relatively high, whereas the lower strength of the rocks at shallow depths limits the size of the earthquakes that can occur here (Scholtz, 1998). Published data for larger UK earthquakes suggest that most events with magnitudes of 4.5 M_w or greater tend to nucleate at depths of at least 10 km or greater. Similarly, although scattered, the relationship between magnitude and depth for UK earthquakes confirms this. Rupture dimensions for the largest recorded earthquakes in the UK are typically of a few kilometres, so although a rupture that nucleates at depth is more likely to propagate into a region of lower strength, the potential for it to reach the surface is limited.

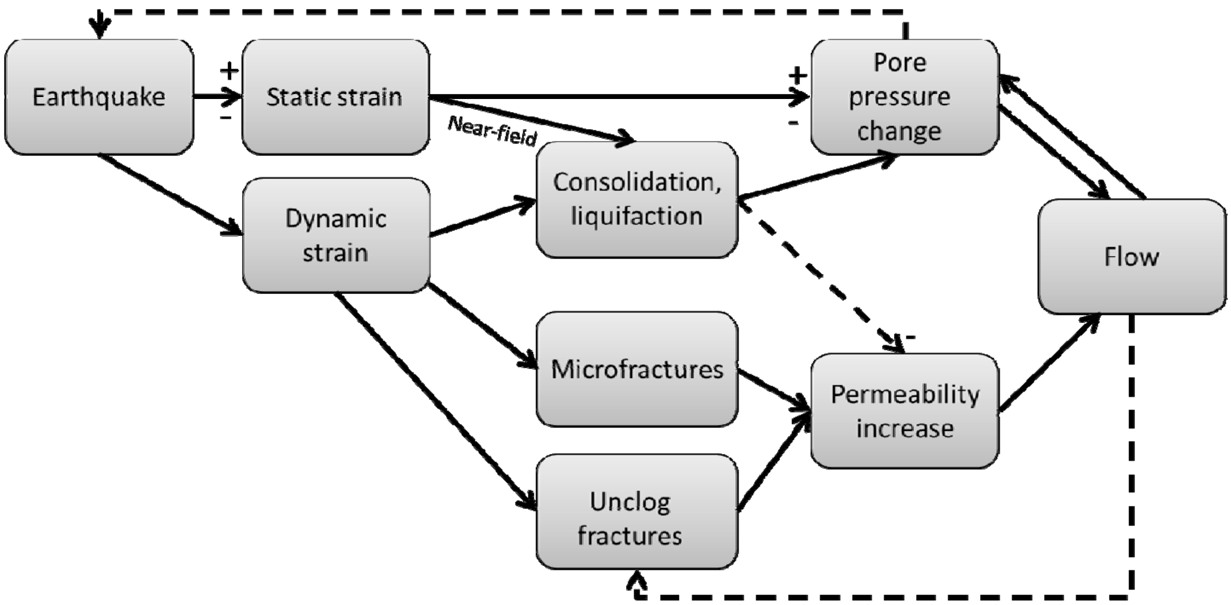


Figure 17: Schematic illustration of the relationship between earthquakes and hydrologic responses; + and - signs indicate the sign of the response (Figure based on Manga and Wang 2007).

Nevertheless, there are a number of examples of large earthquakes in intraplate areas that nucleate in the crystalline basement and rupture through to the surface. For example, the New Madrid earthquakes of 1811 and 1812 in the eastern United States, with estimated magnitudes of 7-7.5 (Hough et al, 2000), ruptured a 100-150 km section of the New Madrid fault zone. However, the magnitudes of these earthquakes are far in excess of the expected maximum magnitudes for earthquakes in and around the British Isles. No British earthquake recorded either historically or instrumentally has produced a surface rupture.

Because earthquakes are likely to occur on pre-existing faults, any underground GDF should be constructed in a location where the number of existing faults is low. However, because this may not be possible, the structure should be able to tolerate expected fault displacements. Fault displacements for larger British earthquakes are likely to be in the range of several centimetres.

3.5.2 Vibration Hazard

A number of studies have examined the differences between ground motion and the surface and at depths of ~100 - ~1000m (e.g. Fukushima et al, 1995; Abercrombie, 1997; De Luca et al, 1998; McGarr and Fletcher, 2005). In general these show that ground motions at depth are less than those at the surface. Differences in ground motions can be explained by:

- The effect of the free surface, which means that, in theory, motion at the surface is twice that at depth;
- Seismic wave velocities of rocks at the surface are less than those at depth, resulting in amplification of ground motions at the surface with respect to those at depth; and,
- Lower attenuation at depth, which can lead to larger ground motions at depths at high frequencies (Abercombie, 1997).

In addition, several studies have documented earthquake damage to underground structures such as tunnels (e.g. Hashash, 2001). These generally conclude that underground structures suffer appreciably less damage than surface structures. This suggests that the shaking hazard for a buried GDF is rather less than that at the surface. A conservative estimate might be a reduction in ground motions by a factor of two, however, this is likely to vary with seismic velocity and attenuation structures. For the Yucca Mountain site in Nevada, USA, (which is no longer under consideration as a GDF site, and which is anyway located in a more seismically active location) Smistad et al., (2001) consider that direct damage from earthquake shaking to the waste packages or engineered barriers of an underground GDF is an insignificant chance, but damage to cladding is a possibility and should be evaluated. Assuming present day seismicity rates, average hazard values of peak ground acceleration for a return period of 10,000 years are in the range 0.1-0.15g (Musson and Winter, 1997).

3.5.3 Secondary Hazards

Earthquake related hydrological changes, such as creation of new springs, changes to water table levels and increased groundwater discharge have all been widely observed after a number of moderate to large earthquakes (Manga and Wang, 2007). These hydrologic responses are a response to strain caused by earthquakes, changing fluid pressures and altering hydrogeological properties such as permeability, which controls the rate of fluid flow. The amplitude of the changes may be large and can often be observed at large distances from the earthquake.

3.5.3.1 Poroelastic Deformation

The presence of fluid in a porous rock modifies its mechanical response, such that the resulting deformation is poroelastic rather than purely elastic. The theory of poroelasticity was first developed by Biot in a series of papers (1941, 1956a, 1956b) and describes how porous solids deform in response to stress, changing pore fluid pressure and allowing fluids to flow.

Fault displacement during an earthquake changes static stress and can result in poroelastic deformation. Water level in a well is a measure of the fluid pressure at depth and changes in water level in wells during and after earthquakes are well documented (e.g. Montgomery and Manga, 2003). Both the sign and magnitude of water level change has been explained by co-seismic static strain generated by the earthquake (e.g. Jonsson et al, 2003), particularly for deep wells in solid rock. This results in rises of water level in zones of contraction and falls in zones of dilatation. Since static stresses decay rapidly with distance, the magnitude of the changes also decays rapidly with distance from the fault, so this is only a significant factor in the near-field.

Seismic waves passing through an aquifer will cause spatial variation in strain resulting in changes in pore pressure and fluid flow. The effect of dynamic strain can cause fluid to flow into and out of wells. Observed water level fluctuations can be several metres for large earthquakes (e.g. Brodsky et al, 2003). These changes have been observed at large distances from earthquakes, however, they decay as a function of time and are unlikely to be a hazard to a GDF.

Muir-Wood and King (1993) use a coseismic elastic strain model to explain increased stream discharge following the magnitude 7.3 Borah Peak earthquake in 1983. They suggest that fluid saturated microcracks open and close in response to the change in stress, causing decreases in porosity and expulsion of deep crustal fluid. However, Rojstaczer et al (1995) suggest that in order to account for the increase in streamflow, a very large volume of the crust would need to be involved, which would require an unrealistic permeability for the crust. Therefore, it seems likely that only comparatively small volumes of water could be released by this mechanism, reducing the potential impact on any GDF.

3.5.3.2 Permanent Deformation

Both static stress and dynamic strain can also cause permanent deformation, which potentially could lead to much larger changes in porosity than poroelastic deformation. Shear deformation of unconsolidated sediments and soils will cause grains to move into pre-existing pore space, reducing porosity. This is known as consolidation. However, in denser deposits this can instead increase porosity (dilatancy). In brittle rocks, application of shear stress above a certain elastic limit causes microcracks to open, increasing rock volume in a process also known as dilatancy (Brace et al, 1966). Higher applied stresses can cause microcracks to coalesce and repeated rupturing along a fault zone can produce significant changes in porosity. As a result, changes in porosity by permanent deformation caused by either static stress or dynamic strain may explain a number of earthquake related hydrological changes.

Large changes in water levels in shallow wells in unconsolidated materials that do not agree with those predicted by co-seismic elastic strain (e.g. Wang et al, 2001; Koizumi et al, 2004) have been explained by permanent changes in permeability and porosity caused by seismic shaking. Similarly, large changes in stream flow have been observed both days and weeks after earthquakes, with excess discharges of 0.7 km^3 observed for the magnitude 7.5 Chi-Chi earthquake, Taiwan (Wang et al, 2001). A number of hydrological models have been used to explain stream flow change including: expulsion of deep crustal fluid (Muir-Wood and King, 1993); changes in near-surface permeability resulting from the formation fractures at shallow depths (Rojstaczer et al, 1995); and, rupture of sub-surface reservoirs. However, most of these models are under-constrained and do not fully explain all observations. In addition, increases in discharge from streams is mainly observed at magnitudes of 6.4 M_w and above, although there are a few documented examples at lower magnitudes, suggesting that the hazard is low, even for the maximum magnitudes expected in the UK. Nevertheless, permanent changes to porosity and permeability resulting from fault rupture and deformation during an earthquake should be considered as potential hazards to a GDF.

3.5.3.3 Liquefaction

Earthquake ground shaking can also cause liquefaction, where the soil loses rigidity and behaves as a viscous liquid rather than as a solid (Seed and Lee, 1966). Liquefaction occurs in sandy soils or unconsolidated sediments that are water saturated, i.e. the space between individual particles is completely filled with water. The earthquake shaking causes the water pressure to increase to the point where the soil particles can readily move. The occurrence of liquefaction depends on two main factors: (i) the soil must be susceptible to liquefaction (Youd, 2003), i.e. loose, water-saturated, sandy soil; and (ii) ground shaking must be strong enough to cause susceptible soils to liquefy.

Liquefaction can change the hydrological properties of unconsolidated near-surface deposits, decreasing porosity and permeability in a way that could affect a GDF. For example, co-seismic consolidation and liquefaction could provide water for observed increases in stream discharge following earthquakes (Manga, 2001). However, as stated above, liquefaction will only occur if the near-surface geology is susceptible and the ground shaking is strong enough. Therefore, the susceptibility should be assessed for any specific site. Liquefaction is generally only observed for earthquakes above a magnitude of 6, however, a number of cases of liquefaction have been observed for earthquakes in the magnitude range 4-6 at distances of several kilometres from the rupture (Wang, 2007). This suggests that the ground shaking from a shallow earthquake of a magnitude that has previously been observed in the UK could cause liquefaction. Liquefaction effects were reported for the 1865 Barrow-in-Furness earthquake, even though, the small magnitude suggests that this is unlikely (Musson, 1998).

3.5.4 Impact of Glaciation Induced Seismicity on a GDF

The possibility of renewed glaciation within the lifetime of a GDF means that estimates of the distribution and rates of regional seismicity cannot be considered to be the same as at present. Geological investigations in a number of regions have found evidence for significant post-glacial movement of large neotectonic fault systems, which were likely to have produced large earthquakes around the endglacial period. For example, Lagerbäck (1979) suggests that the 150 km long, 13 m high fault scarp of the Pårve Fault in Sweden was caused by a series of post-glacial earthquakes. Adams (1989b, 1996) finds evidence for post-glacial thrust faults in eastern Canada. Muir-Wood (2000) suggests that the large historical earthquakes of the New Madrid sequence (1811-1812) and Charleston (1886) in North America may have been due to an outward moving wave of forebulge collapse from the Laurentian ice sheet. However, Wu and Johnston (2000) argue that, because rebound stress decays away from the former ice margin, glacial unloading is unlikely to have triggered the New Madrid earthquakes. Davenport et al (1989) and Ringrose et al (1991) have found similar evidence for post-glacial fault displacements in Scotland. However, Frith and Stewart (2000) argue that these are restricted to metre-scale vertical movements along pre-existing faults.

The primary component of deformation in response to the loading and unloading of ice sheets is GIA. Measurement and observations of sea-level change have been commonly used to constrain models of GIA (e.g. Tushingham and Peltier 1991; Mitrovica 1996; Lambeck et al 1998) that include the effects Earth rotation as well as three dimensional effects of the ice sheet. Numerical models have also incorporated 3D Earth models that include gravitational effects (e.g. Kaufmann et al 2000). The GIA of British Isles has been studied by a number of authors (e.g. Milne et al., 2006; Bradley et al 2009). These models are constrained by a high quality sea-level dataset (Shennan and Horton, 2002).

Some of the current understanding of the influence of glaciation on seismicity is summarised by Stewart et al. (2000). Early work by Stein et al (1979) suggested that rebound stresses alone could cause earthquake failure. Quinlan (1984) suggested that the rebound stresses probably act as a trigger for faults that are already close to failure. Wu and Johnston (2000), among others, examine differences in fault stability margins to estimate if faults are moved closer to failure by the deformation caused by an ice sheet. In short, a number of studies (e.g. Pascal et al, 2010) suggest that earthquake activity beneath an ice sheet is likely to be suppressed, followed by much higher levels of activity after the ice has retreated again. Consequently, estimates of seismicity based on current rates may be quite misleading as to the possible levels of activity that could occur in the more distant future. It should be noted that the largest stress changes occur at the former ice margins, making these the most likely source region for enhanced earthquake activity. The implication for a GDF in such a region is that seismicity rates following any future glacial period may be significantly higher than at present. Given our current maximum magnitude in the UK of around 6 it is not unreasonable to expect an increase in the maximum possible magnitude to 7 following such an event. However, it should be noted that post-glacial fault stability is dependent on not only the thickness and extent of the ice sheet, but also on the initial state of stress and the properties of the Earth itself, such as stiffness, viscosity and density (Lund, 2005).

3.5.4.1 IMPACT OF TSUNAMI

Tsunami, sometimes loosely referred to as 'tidal waves', will not affect a closed GDF but it is possible that the surface facilities associated with an active GDF sited on the coast may be affected by tsunami so these phenomena are briefly discussed in this section.

Tsunami are typically generated by vertical uplift of the seafloor over a large area during earthquakes. In general, only earthquakes with magnitudes in excess of 7.5 M_w are capable of generating tsunami and the main source regions for such events are at plate boundary subduction zones where one of the Earth's tectonic plates is being pushed down, or subducted, beneath another. Places where this happens include Japan, Sumatra and South America all of which have had earthquakes of magnitude 8.5 or greater in the last few years that have resulted in tsunami. By contrast, the British Isles sits in the middle of a tectonic plate, Eurasia. Our nearest plate boundary is at the mid-Atlantic ridge, where the earthquakes are too small to generate tsunami. The nearest subduction zones to Britain lie at the Hellenic Arc, south of Greece and in the Caribbean. Tsunami have occurred in both these regions in historic times, but did not affect the UK. The largest recorded British earthquake had a magnitude of 5.5 M_w and was over 65,000 times smaller than the Tohoku earthquake in Japan (Musson, 1994). Although it occurred under the North Sea it was too small to generate a tsunami. This event is close to the maximum credible magnitude for a British earthquake.

However, tsunami have impacted on the British Isles in the past. Over 7000 years ago, a massive submarine slide off the coast of Norway, known as the Storegga slide (Bondevik et al, 1997), resulted in a tsunami reaching the northeast coast of Britain. Evidence of this can be found in geological deposits from northeast England to north of the Arctic Circle. These show that the wave reached over 20 m above sea-level at Sullom Voe, Shetland. However this quickly decreases to the south with 3-4 m in northeast Scotland and 1 m in northeast England. A repeat of this event is unlikely, since geological models suggest that another Ice Age is needed to re-establish the conditions for a similar failure.

In 1755, Lisbon was destroyed by a magnitude 8+ earthquake and tsunami. The tsunami reached the southwest coast of England and its arrival in Mount's Bay, Cornwall was observed by the naturalist William Borlase (1755), who described several large waves crashing against the shore over a period of two hours. A study by BGS, HR Wallingford and the Proudman Oceanographic Laboratory commissioned by Defra looked in detail at the possible effects of another earthquake like the 1755 Lisbon earthquake (Richardson et al, 2006). Modelling results suggest that the wave would take around 5 hours to reach Britain, with maximum wave heights of 1-2 m around the majority of Cornwall. Such wave heights are similar to those resulting from typical storm surges that are experienced on a far more frequent basis.

3.5.5 Impact on Groundwater

Seismic events are likely to have only transient impacts on groundwater flow. As noted above, seismic waves from an earthquake can cause groundwater to flow into and/or out of the rock mass. This may include increased discharge into streams over a period of several weeks following a large earthquake. Earthquakes of the magnitude likely to be experienced in the UK are unlikely to have any effect on groundwater flows. Permanent deformation may lead to increase or decrease in bulk rock permeability, mainly in the vicinity of the active fault.

4 Climate change and glaciation

Changes in climate over the next million years have the potential to have a significant influence on processes that drive radionuclide migration and associated biosphere impacts. This is caused by their influence on eustatic sea-level and hydrological systems, and also by governing the rate and form of erosion and sediment dispersal from the land surface. In this context, the future timing, intensity and frequency of glacial and interglacial episodes will be fundamental in determining the long term evolution of the biosphere and geosphere both regionally and globally. Future glacial-interglacial cyclicity will directly influence hydrology and hydrogeology, as well as the rates and styles of erosion and sediment transport across the UK landmass. The depth of groundwater recharge will be increased during periods of glacial melting and river courses will increase in length as they drain towards lowered base levels Landscape modification will range from glacial erosion and sediment dispersal, by topographically controlled glaciers, and by topographically enveloping ice sheets during glacial episodes, to fluvially-dominated incision and slope modification during temperate periods.

The nature, thickness and extent of ice cover across the UK, as well as the duration of each glaciation, will control levels of isostatic depression of the land surface. These factors associated with ice sheet loading and uplift because of isostatic rebound, will also directly influence near-surface stress fields within the shallow geosphere, control regional relative sea-level, the hydrological base level and indirectly change groundwater regimes. The extent and depth of permafrost, developed under periglacial conditions will alter the geomechanical properties of the land surface, reduce the recharge of aquifers by precipitation and alter the positions of recharge and discharge locations, relative to those that prevail under present day temperate conditions. Many of these effects, most notably eustatic sea-level, would also be influenced by changes in global ice volume due to ice sheet growth in areas remote from a GDF during periods of global cooling. Climate change and changes to sea-level in near coastal areas will also impact on groundwater in terms of, for example, composition, flow pathways and flow rates.

In previous biosphere and geosphere reports, (NDA, 2010b; SERCO, 2011), assessments of changes in biosphere and geosphere systems during the evolution of a GDF have adopted a methodology based on the identification of Features, Events and Processes (FEPs) causing environmental change. FEP analysis for this purpose has been widely endorsed by international organisations such as the International Atomic Energy Agency (IAEA) and the Nuclear Energy Agency of the Organisation for Economic Co-operation and Development (NEA/OECD). FEPs can be divided into those that are external, or internal to the disposal system and its geographical setting. Internal FEPs characterise the disposal system, whilst external FEPs impose change upon the system. In this report we concentrate on the latter. A schematic illustration of external FEPs and their influence on the biosphere and geosphere is given in Figure 18. It illustrates the major roles played by global climate change and regional climate regime and their interactions with the other processes and events affecting the natural evolution of the geosphere and the related changes to the biosphere of the UK.

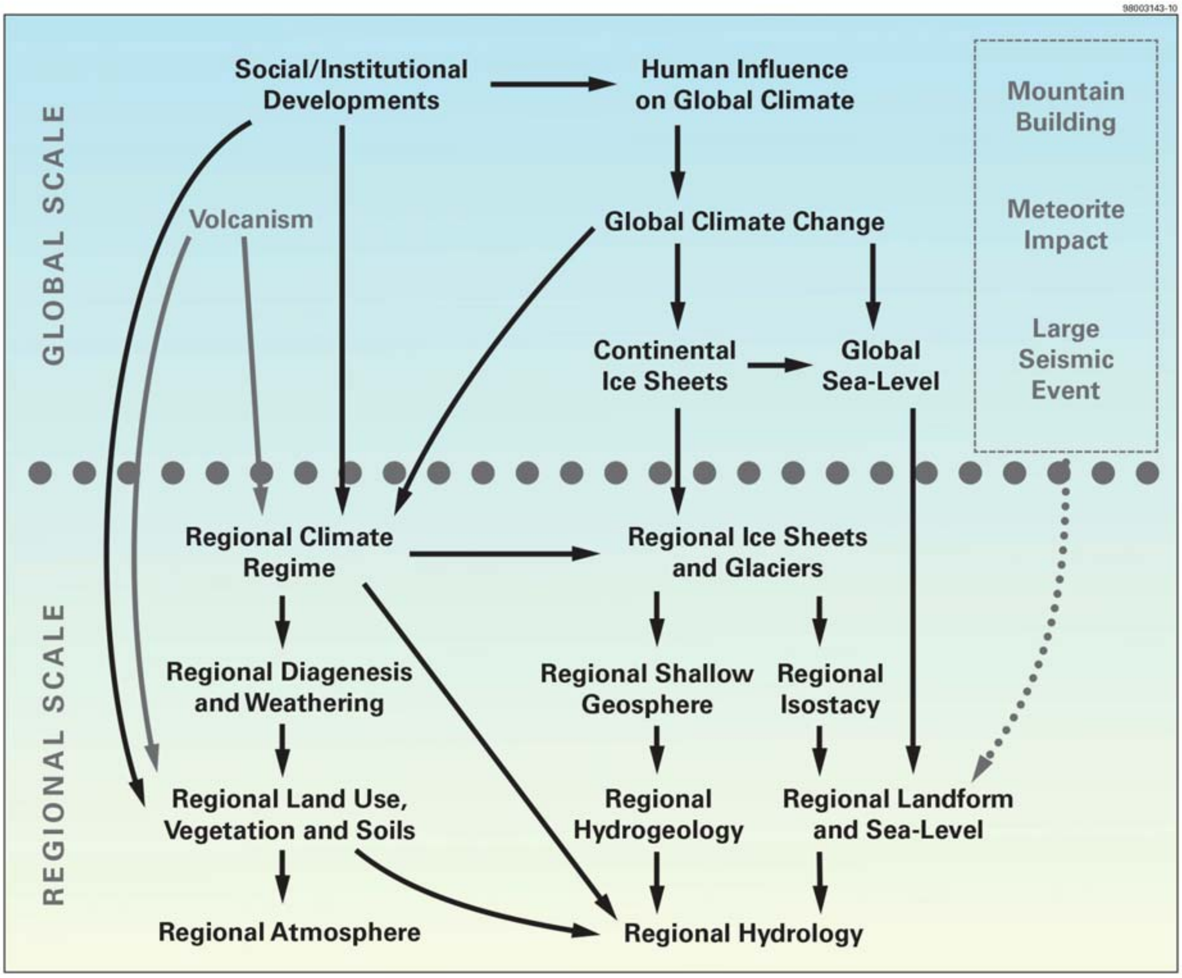


Figure 18. Illustration of the external FEP model for describing the influence of change on the biosphere and geosphere systems. Figure modified after NDA (2010b). External features events and processes in grey text are screened out from consideration.

4.1 DEFINING CLIMATE SCENARIOS

This section contains a review of current published research and modelling of the nature of past and future climate change, concentrating on the frequency and duration of glacial and interglacial episodes and assessing their intensity relative to those documented from the Quaternary record (in particular those from the last 800 ka). Projected glacial-interglacial cyclicity, driven by 'natural' orbital and CO_2 forcing, and the expected duration and degree of understanding of influence of anthropogenic warming on this natural cycle, have been assessed.

Whilst the focus of this report is on natural change, the impact of human induced climate change, caused principally by the release of CO_2 from combustion of fossil fuels, is also covered in this chapter. Nevertheless, assessment of the nature and extent of the climate change resulting from anthropogenically induced greenhouse gas emissions remains challenging and a source of considerable uncertainty. Full consideration is recommended of how the trends identified in this report marry with those illustrated by a wide range of reports, including the 2007 Intergovernmental Panel on Climate Change Fourth Assessment Report (Solomon et al, 2007) (http://www.ipcc.ch/) including uncertainties associated with the modelling of these, for instance being explored by the Climate Model Intercomparison Project (http://cmip-pcmdi.gov)

4.1.1 Glacial – Interglacial cycles

Over long periods of time the temperature of the Earth alternates between cold, glacial phases and warm interglacial phases. Minor changes in the Earth's orbit can result in a 4 ºC change in the global mean temperature, which has a very dramatic impact on the Earth system (Cedercreutz, 2004). These changes were first defined mathematically in the Milankovitch theory, which describes the collective effects of changes in the Earth's orbit upon its climate, variations in precession (wobble), tilt and eccentricity, of the Earth's axis determined climatic patterns through orbital forcing. The periodicities of these 'Milankovich Cycles' are approximately 22 ka, 41 kyr and 100 kyr respectively; findings subsequently verified by investigation of the δO^{18} composition of benthonic foraminifera microfossils recovered from deep-ocean cores (Hays et al, 1976; Imbrie et al, 1984) and by measurements of deuterium (D) from Antarctic ice cores (Petit et al, 1999). The semi-analytic planetary *Variations Séculaires des Orbites Planétaires* (VSOP) theory enables prediction of simplified Milankovich cycles for at least the next 800 kyr (Bretagnon and Francou, 1988).

During the Cenozoic Era (the last 65 Ma) glaciations began in Antarctica during the mid-Eocene (35-40 Ma) (Ingólfsson, 2004), but records from cores in Lake Baikal, suggest that a turning point in climate towards colder conditions in the Northern Hemisphere occurred about 4 Myr (Kashiwaya et al., 2003). However, the consensus is that global cooling generally occurred later, at 2.5–2.8 Myr BP, (Mangerud et al., 1996; Haug et al., 2004); at the beginning of the Quaternary Period. Apart from the Greenland ice sheet, which formed about 18 Myr (Thiede et al., 2011), the first large ice masses in the Northern Hemisphere, developed about one million years ago (Muller and MacDonald, 1997).

The initial 4 Myr cooling has been attributed to the slow changes in the global configuration of the continents as a consequence of sea-floor spreading. These include the emergence of the Panama Isthmus and the deepening of the Bering Straits, both of which had pronounced effects on the ocean circulation patterns (Raymo, 1994). The cooling has also been attributed to the uplift of high mountain ranges such as the Tibetan Himalayas and the Sierra Nevadan and Coloradan mountains of North America, causing perturbations of the upper atmosphere and subsequent climatic changes (Ruddiman and Kutzbach, 1991). Uplift of the Himalayas also may have resulted in a massive increase in chemical weathering during the late Cenozoic, leading to increased sedimentation of calcium carbonate and atmospheric depletion of carbon dioxide. Global cooling, the inverse of the 'greenhouse effect' would thus have occurred (Raymo, 1994; Raymo and Ruddiman, 1992). However, none of these mechanisms can wholly explain the rapid intensification of glaciation observed in the deep ocean record at about 2.8 to 2.5 Myr (Maslin et al., 1998).

Most researchers have suggested that, since the 'Mid-Pleistocene Revolution' (MPR) c. 0.9-1 Myr BP, the eccentricity of the Earth's orbit has been the dominant orbital 'forcing' mechanism of global climate and glaciations have followed this 100 kyr cycle (Mangerud et al., 1996; Raymo, 1994; Ruddiman and Kutzbach, 1991; Ruddiman, 2003a). Before this time, glaciations followed the 41 kyr cycle. This shift of the dominant glacial cycle from 41 kyr to 100 kyr has been attributed to many causes, including an increase in the amount of interplanetary dust (Muller and MacDonald, 1997). It has also been attributed to tectonic movements, changes in atmospheric CO_2 and other 'greenhouse' gases, as well as to fluctuations in primary solar insolation (Kashiwaya et al., 2003; Ruddiman, 2003a). The fluctuations were also amplified substantially by additional factors involving physical, biological and chemical interactions and 'feedback loops' between the atmosphere, oceans and ice sheets. For example, complex changes in the surface and deep water circulation patterns of the ocean, are said to have played a crucial role (Broecker and Denton, 1990).

Some research has questioned the viability of eccentricity of the Earth's orbit as the dominant forcing mechanism of global climate change after the MPR (Maslin and Ridgwell, 2005) because it has by far the weakest influence on insolation received at the Earth's surface of any of the orbital parameters. This research has also sought to explain the delay between the significant increase in global ice volume, which occurred around the time of the MPR and the amplitude of the 100 kyr cycle which abruptly increased much later, around 650 ka. The alternative explanation of the transition from 41 kyr to 100 kyr glacial interglacial cycles is that it is more closely linked to a combination of cycles of precession of the Earth's axis and of its orbit, which are determined by the eccentricity of the Earth's orbit, but are not driven by it.

Of particular importance to the climate of the British Isles are changes in the position and strength of the Gulf Stream. This northward-flowing current of warm surface water is compensated by the return southwards of cold, dense saline water at depth. Sudden changes in this circulation pattern, which is the northern part of the global Meridional Overturning Circulation (MOC); also known as the 'North Atlantic Conveyor', may have had a major impact on regional climate (Skinner and Porter, 1995). For example, large volumes of fresh water released during rapid warming events within the whole Atlantic area, during the decay of the last Northern Hemisphere glaciation may have temporarily 'switched off the Conveyor', by diluting its salinity. This would have led to contemporaneous localised cooling in north-western Europe (Lagerklint and Wright, 1999). Modelling the operation of the MOC under present conditions (Figure 18), at the Last Glacial Maximum (LGM[1]) ~ between 33.0 and 26.5 kyr BP, and during a subsequent release of meltwater, have all indicated that the operation of the conveyor was substantially weaker during the LGM than at present (Seidov and Haupt, 1997). The simulations indicated that major changes occurred (including reversal of the Indian-Atlantic branch of the MOC due to interruption of North Atlantic Deep Water production) because of very localized meltwater input.

The studies of isotope geochemistry and micropalaeontology of deep ocean-floor sediments, which validated the general principles of the Milankovich theory, have also revealed that global climatic conditions have fluctuated continuously throughout the Quaternary period and that the climate system switched rapidly between *interglacial* and *glacial* modes. Although the precise timing and duration postulated for each cycle is a matter of on-going research and debate, at least 50 significant 'cold–warm–cold' oscillations have been recognised (Merritt et al., 2003). The 'SPECMAP' (SPECtral MAping Project – funded by the US NSF) deep ocean oxygen isotope record indicates that, up until about 760 ka, each dominant warm–cold cycle lasted about 40 kyr (Ruddiman et al., 1989). During the cold phase of each cycle ice caps in Greenland, Alaska, Iceland and Scandinavia expanded to the coast, and glaciers probably developed in the western Scottish Highlands at high elevations (Clapperton: 1997). These events were probably slightly diachronous, the resolution of their timing is still a matter of active research and is a function of the quality and number of proxies that have been studied.

To date there have been seven major glacial–interglacial cycles (Figure 19). Each 'glacial' episode lasted between 80 and 120 kyr and was followed abruptly by a warm interglacial

1 The Last Glacial Maximum (LGM) is conventionally defined from sea-level records as the most recent interval in Earth history when global ice sheets reached their maximum integrated volume. The term has also been used by many authors for the time(s) when individual ice sheets in the northern hemisphere each reached their maximum extent. These were diachronous events, so that the time span for the LGM, for Fennoscandia, for example, is often quoted as c. 18 ka BP.

interval lasting 10 to 15 ka. The rapid deglaciations are defined as ‘Terminations I–VII’ (Broecker, 1994). The glacial periods included long, cold intervals (stadials), and less cold, and even warm, intervals lasting for a few thousand years called (interstadials).

Abrupt deglacial events during the last 60 kyr were associated with ‘Heinrich events’, which accompanied periodic destruction of northern hemisphere ice shelves, and the consequent release of a prodigious volume of sea ice and armadas of icebergs into the North Atlantic. Melting icebergs deposited rock debris that had been eroded by the continental glaciers from which they originated. This debris was dropped onto the sea floor as "ice rafted debris" (IRD), which is now found in deep ocean sediment cores. It was principally derived from the Laurentide ice sheet, but ice sheets in Iceland, Britain and Scandinavia are now known to have contributed IRD during several Heinrich events (Sejrup et al., 2000; Walden et al., 2006).

Six principal Heinrich events (H1-6) are generally recognised, although some authors (Broecker, 1994; Bond and Lotti, 1995) associate the last (Younger Dryas) glaciations in Europe (Loch Lomond Stadial glaciations in Britain) with a seventh Heinrich Event (H0). An earlier (undated) H11 was also recognized in Heinrich’s original study (Heinrich, 1988). The precise causes, timing and duration of Heinrich events is still a matter of active scientific debate. It is generally agreed that Heinrich events are rapid. Some estimate that they can last for around 200 to 2,300 years, with an average duration of 500 years (Hemming, 2004); others estimate an average duration of 750 years, and note that their abrupt onset may occur within a few years (Maslin et al., 2001). Most, but not all, Heinrich events precede rapid warming events known as Dansgaard–Oeschger (D-O) events, which are repeated around every 1,500 years (Bond et al, 1999) and are recognized from oscillations in the $\delta^{18}O$ signal in the Greenland Ice Core Project (GRIP) and the second Greenland Ice Sheet Project (GISP2) ice cores. Different authors give slightly different timings of the last four Heinrich events:

H0: c. 12 kyr BP;

H1: 14 kyr BP or 16.8 kyr BP;

H2: 22 kyr BP, 23 kyr BP or 24 kyr BP;

H3: c. 31 kyr BP to 29 kyr BP;

H4: 35 kyr BP, 37 kyr BP or 38 kyr BP;

most agree a date of 45 kyr BP for H5; and

c. 60 kyr BP for H6.

This variation is largely a function of the quality of the proxies in the sediment cores and the accuracy of the dating methods that can be employed (H1 and H2 are commonly calibrated by ^{14}C dating; older events by correlation with the GISP2 ice core record).

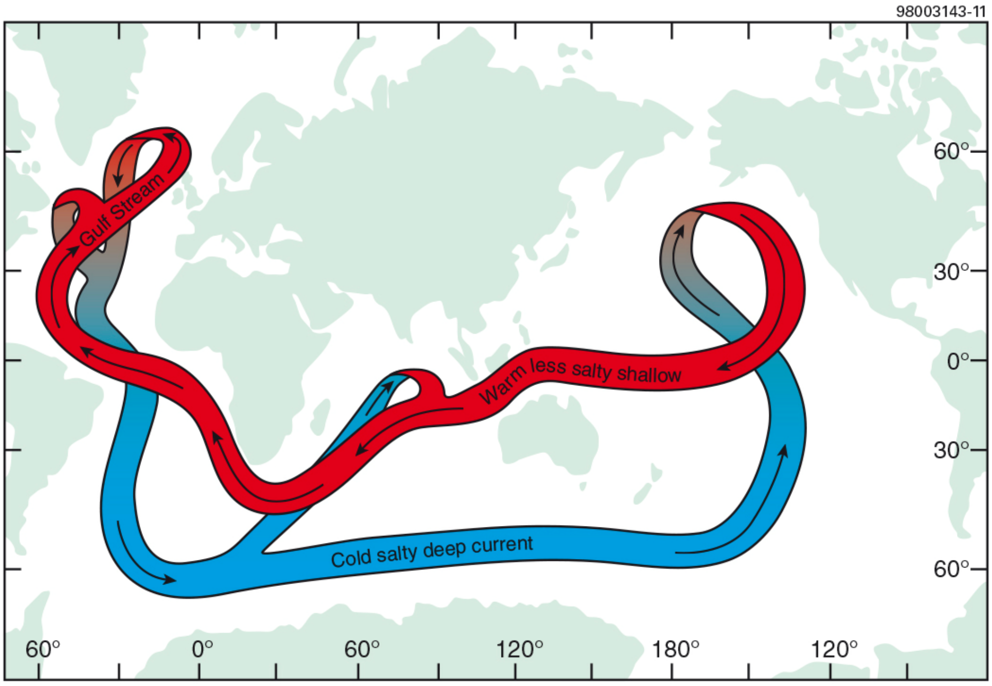


Figure 18. Major thermohaline cells of the MOC, (modified after Skinner and Porter (1995)). The cells are driven by exchange of heat and moisture between the atmosphere and the ocean.

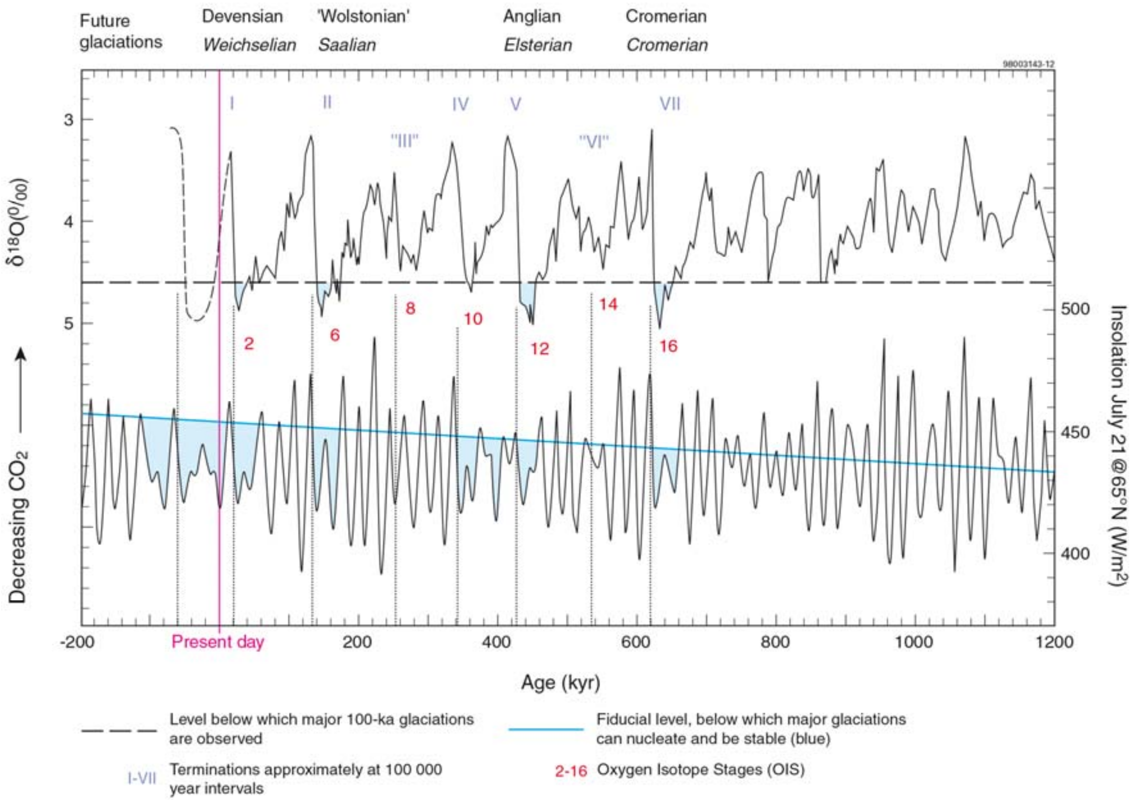


Figure 19 Oxygen isotope curve (after Raymo (1997) from Pacific site 849, representing ice volume change over the last 1.2 Myr and a comparable summer insolation curve for 65°N (modified after Merritt et al (2003)).

Various mechanisms have been proposed for the initiation of Heinrich events; in particular instability of major ice sheets, the so called ‘binge-purge model’ whereby increase in mass of the ice sheet past a critical point leads to rapid extension of the ice front by basal sliding and its eventual collapse by thinning and iceberg calving (MacAyeal, 1993). Others have suggested that changes in the flux of solar energy, which occur on about a 1,500 year periodicity during the present Holocene interglacial and may extend back into the last glacial period, correlate to Daansgard-Oeschger cycles, and be related to Heinrich event initiation (Bond et al., 1997). These ‘Bond cycles’ have been recognized by IRD from Holocene sequences in ocean cores. Other proposed mechanisms include sea-level rise by thermal expansion due to climate warming, leading to floating of formerly grounded ice-margins and increased iceberg calving; recent work suggests that triggering of the last Heinrich event ascribed to the Laurentide ice sheet was caused by slowing and warming of the Atlantic meridional overturning circulation (Alvarez-Solas and Ramstein, 2011). This occurred about 1000 years before the H1; it increased basal melting of the Labrador ice shelf and led to its break up.

The foregoing discussion on the mechanisms, lags and feedbacks of global and local climate, ice sheet dynamics and oceanic and atmospheric systems on past ice sheet growth and decay, illustrates the degree of complexity and uncertainty of the understanding involved in modelling past climate states. The general principles are broadly understood, but precise details of the frequency and duration of past glacial, interstadial and interglacial episodes are still a matter of active research. The resolution of past climates during the historical past (since written records began) increases as modern instrumental and written records become available, but resolution decreases in pre-history as one passes further and further back in time. Time constraints on proxy records from terrestrial and marine sequences generally become more lax and subject to error, once the threshold for ^{14}C dating is passed (at the most c. 50 kyr BP). Because ^{14}C decay is logarithmic, the ratio of ^{14}C halves about every 5,730 years (the ^{14}C half-life). Consequently, in very old material (regardless of the size and purity of the sample), beyond almost 9 half-lives, insufficient ^{14}C will be present to enable accurate age determinations to be made. Luminescence and amino acid dating, however, can push the chronological control back hundreds of thousands of years, given suitable material to analyse. Cosmogenic isotope dating of rock surfaces and erratic boulders can also span the gap between ^{14}C, and other dating methods; in suitable circumstances it can be used to date the former positions of ice margins and ice limits from beyond the last glacial cycle. In ice cores from Antarctica, apparently continuous records of climate change covering the last four glacial-cycles (420 kyr BP) have been constructed based on Deuterium $\delta D^{0/00}$ (Petit et al., 1999) with respect to Standard Mean Ocean Water (SMOW), calibrated in places by dust records and by radiometric dates on volcanic ash horizons. However, as depth down the core increases, these gases, lithic and other records become increasingly tuned by other proxy data sets from ocean cores and by graphical statistical fits to Milankovich cycles.

4.2 DEFINING FUTURE CLIMATE STATES

Because climate has such a fundamental role in the natural evolution of the geosphere and biosphere of Britain, it is useful to try to establish the timing and duration of future climate states over that time span. In previous studies (NDA, 2010b; Serco, 2011) a small number of future climate states that are believed to be applicable to the future climate evolution of Britain, were defined in terms of monthly temperature and precipitation (Table 2). The modelling of future climates was then addressed in terms of the expected timing, duration and the nature of transitions between these climate states under a variety of future climate scenarios.

Climate Group	Monthly Temperatures	Climates	Notes
Sub-tropical	Over 17 ºC in all months	Subtropical rain Subtropical summer rain Subtropical winter rain	Mediterranean-type climates
Temperate	Over 9 ºC in 8 to 12 months	Temperate oceanic Temperate continental	Present-day UK climate
Sub-arctic	Over 9 ºC in 1 to 3 months	Subarctic oceanic Subarctic continental	Boreal Periglacial forest tundra
Polar	Over 9 ºC no months	Tundra Ice	Periglacial (permafrost) Full glacial

Table 2. Classification scheme used to describe climate states. Table based on (Rudlof, 1981).

A similar approach has been adopted for assessing future climate scenarios at potential GDF sites in Fennoscandia (Cedercreutz, 2004; SKB, 2006b; SKB 2011). It can also be used to describe, in general terms, the past climate of the British Isles since the MPR.

4.3 SUMMARY OF THE PROXY EVIDENCE FOR PAST CLIMATE STATES FOR THE BRITISH ISLES DURING THE LAST 800KYR

The terrestrial sedimentary record of past climate change in Britain since the MPR is fragmentary, because proxy evidence of early events has largely been removed by subsequent glacial episodes and even Post-glacial (Holocene) sequences contain gaps caused by erosion and non-deposition. The most complete records come from deep ocean cores, such as that from Ocean Drilling Program (ODP) site 677, which can contain sedimentary sequences spanning the whole of the Quaternary and can be calibrated by the $\delta\ O^{18}$ and geomagnetic polarity records (Figure 20a) (Shackleton et al., 1990). The terrestrial record of northern hemisphere glaciations can also be calibrated by $\delta\ O^{18}$ records from ice cores, such as that from the GRIP Summit ice core (Figure 20b). The $\delta\ O^{18}$ record is divided into Oxygen Isotope Stages (OIS), which are numbered sequentially; odd numbers indicate global warm periods and even numbers indicate global cold periods. Most warm periods equate with interglacial or interstadial episodes and cold periods with glacial episodes, but for some of the cold episodes prior to OIS 5, which began c. 128 kyr BP, the records of terrestrial glaciations in north-west Europe and Britain are very poorly understood or controversial. In Britain, in particular, this applies to glaciations in OIS 8, 10, 14, 16 and 18; earlier episodes are even less well understood (see Figure 20).

Since the onset of 100 kyr cyclicity each global ‘glacial’ episode has lasted about 80-120 kyr and was followed abruptly by a warm interval lasting 10 to 15 ka. Clearly, even this simple pattern does not mirror directly the Milankovich cyclicity. Because of the lags and leads provided by complex land, ocean atmosphere interactions alluded to in Section 4.1.1, timing of the onset, and the duration of glacial, interstadial and interglacial conditions in Britain and elsewhere in north-west Europe was diachronous. Areas such as Fennoscandia, which are at higher latitudes, have been subjected to more intense glaciations, more frequently than Britain. Since 600 kyr BP, for example 8 glacial (OIS 2-3-4, 3, 5b, 5d, 6, 8, 12), 3 major interstadial (OIS 5a, 5c, 7) and 3 interglacial episodes (OIS 1, 5e, 11) have been recognised in Finland (Cedercreutz, 2004).

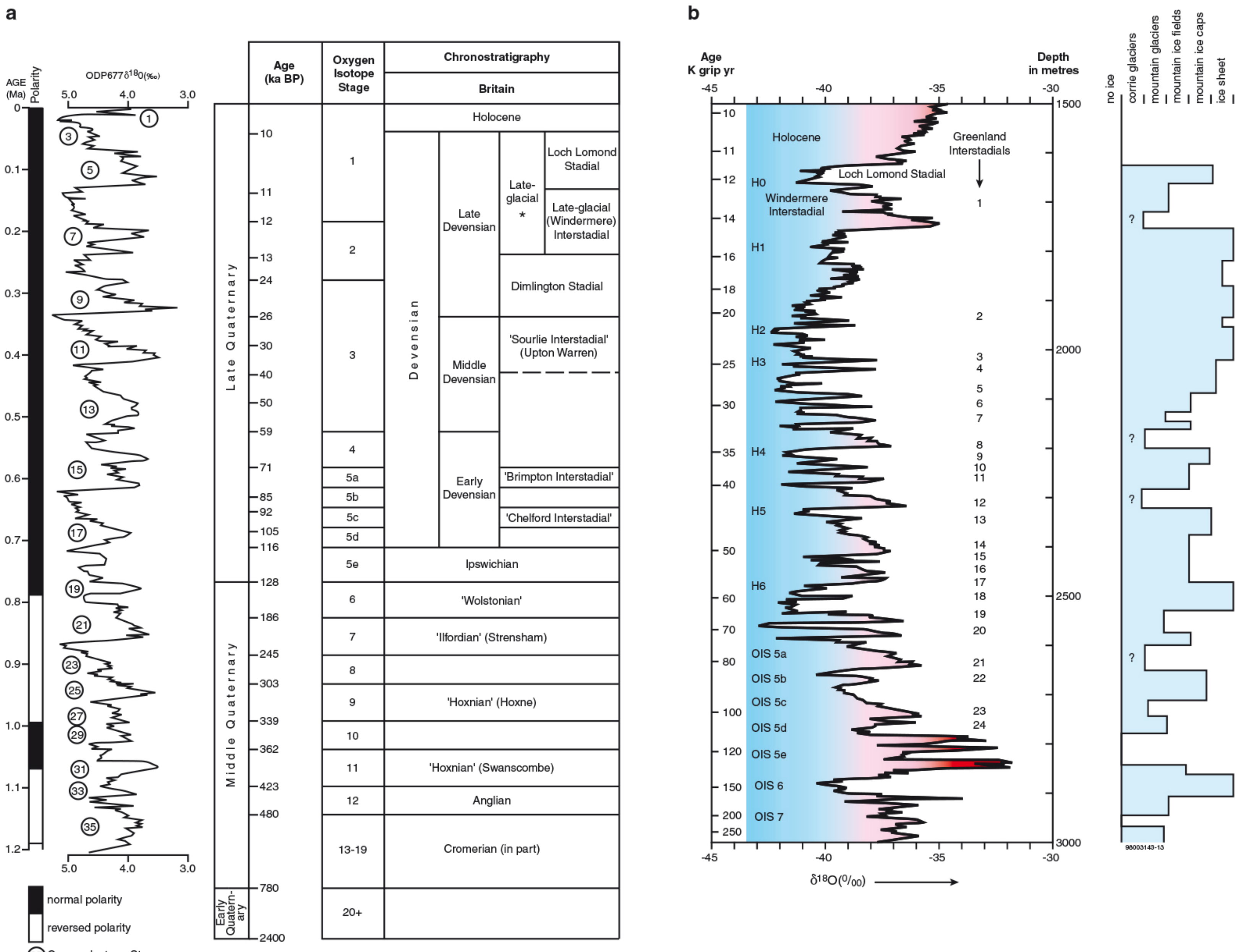
Ice sheet modelling of the last 100 kyr for NW Europe, using the University of Maine Ice Sheet Model (UMISM) (Näslund et al., 2003; SKB, 2006b) indicates restricted glaciations in the south western mountains of Norway in OIS 5d, an ice free period in OIS 5c, extensive glaciation of the Swedish and Norwegian uplands and Scotland in OIS 5b, an ice free period in OIS 5a, extensive glaciation of Baltoscandia and Britain by confluent ice sheets in OIS 4, restricted glaciation of the Swedish and Norwegian uplands in OIS 3 extensive glaciation of Baltoscandia and Britain by confluent ice sheets in OIS 2 and restricted glaciation of Baltoscandia and Scotland in early OIS 1 (the Younger Dryas Stadial of north-west Europe and the Loch Lomond Stadial of Britain).

4.3.1 Offshore Evidence

The offshore record from the northern portion of the British continental shelf tentatively indicates that regional glaciation first impacted on the northern North Sea before 1 Myr BP (Merritt et al., 2003) and several glacial episodes are thought to have occurred prior to 780 kyr BP. The oldest known glacial deposits laid down by ice flowing from the Scottish Highlands have been attributed to glaciation in OIS 18, within the Cromerian Complex; late Cromerian interstadial sediments, overlain by arctic glaciomarine sediments thought to be of Elsterian (OIS 12) age are also present. Tunnel valleys and some of the sediments within them have been correlated with the Elsterian of north-west Europe, whilst channelled surfaces at the top of the Aberdeen Ground Formation probably formed in more than one glacial cycle, including a severe Cromerian glaciation during OIS 16 (Holmes, 1997). Interglacial deposits assigned to OIS 11 and OIS 9 overlie a widespread erosion surface assigned to the Elsterian glacial event. This indicates that the British and north-west European ice sheets may have been confluent in OIS 12 and that much of the continental shelf would have been glaciated at that time.

Glaciomarine sediments, and glacial till occurring beneath marine intervals containing foraminifera assemblages typical of the Eemian/Ipswichian Interglacial (OIS 5e) suggest that there was a regional glaciation of the North Sea basin during OIS 6 and possibly an earlier more limited OIS 8 glaciation in the northern part of the basin. The OIS 6 glaciation is thought to have affected most of Scotland, because ice flowing from the mainland affected the northern North Sea at least as far south as 56°N (Sutherland and Gordon, 1993) and it reached the shelf break to the north-west of Scotland. Some workers have also concluded that ice streams occupied the Norwegian Channel during every glacial stage between OIS 12 and 6; each one representing a separate regional glaciation (Sejrup et al., 2000).

Figure 20. (page 66) British Quaternary chronostratigraphy and a representative oxygen isotope and geomagnetic polarity record (ODP 677) (modified after Merritt et al (2003)); b. Greenland (GRIP Summit ice core) oxygen isotope record. The GRIP timescale was determined by counting annual ice layers back to 14.5 kyr BP; and beyond this time by estimation based on ice flow modelling.



* Note: The 'Late-glacial' spans the time between the retreat of the Main Late Devensian ice sheet and the beginning of the Holocene.

Micromorphological examination of sediment cores suggested that a minimum of three major glacial episodes have occurred in the North Sea Basin between OIS 4 and OIS 2, and that the British and Scandinavian ice sheets were confluent in the central North Sea on at least two occasions (Carr et al., 2006). This is a conclusion supported by the UMISM modelling cited above and by more detailed modelling of a confluence zone that might be applicable to the growth and decay of the Fennoscandian (FIS) and British ice sheets on more than one occasion (Boulton and Hagdorn, 2006).

4.3.2 Onshore Evidence

Onshore evidence in Britain for climate states during the Quaternary is widespread, detailed and in some cases subject to considerable scientific debate. As is the case with climate cyclicity, the general parameters are clearly understood, but precise details of, for example the south western limit of the OIS 2 ice sheet, are still a matter of active research. Because the distribution of earlier sediments is fragmentary within the limits of the last glaciation, records of previous glacial, interstadial and interglacial intervals are most abundant in the south eastern portion of Britain and within the alluvial sequences of the drainage basin of the ancestral River Thames, which was largely unglaciated at this time.

Pre Anglian Glaciations

Evidence of the earliest glaciation in the North Sea basin is provided by far travelled, fresh, heavy minerals and small erratic clasts of chert from Yorkshire that occur in blue, laminated sublittoral marine clays, at Easton Bavents on the Suffolk coast. It is also indicated by fossil assemblages from similar deposits at nearby Covehithe. The source and character of the minerals and rocks suggest that they were transported southwards into the region by a 'North Sea ice sheet' of unknown extent, and then redistributed by ice rafting (Rose et al., 2001). Support for this Baventian (OIS 68-72) glaciation is also provided by analysis of pollen, foraminifera and molluscs, which indicate arctic marine conditions, adjacent to a land mass with an interstadial climate and a vegetation cover of grass and heath, and woodland in sheltered localities (West et al., 1980).

Later early Quaternary glaciations of Britain are indicated by the presence of far travelled erratic clasts, from glaciers developed in North Wales and the West Midlands, in terraced river gravel sequences in East Anglia and the Thames valley. Many of these gravel deposits accumulated in cold interstadial (periglacial) conditions, as suggested by the presence of intra formational ice-wedge casts, in several of the sequences (Bowen et al., 1986). Fossil faunas and floras, indicating both cold and temperate climates have also been described from organic channel infills, and palaeosols indicating both temperate and arctic conditions have been recorded on the surface of many of the gravels (Rose et al., 1999).

The first clear onshore indications of glacial climate states in Britain since 800 kyr BP occur within sediments of the ‘Cromerian Complex’ exposed in coastal cliffs in East Anglia. As its name suggests, the stratigraphical subdivision of this interval is difficult, and a subject of continuing research (Preece, 2001; Preece and Parfitt, 2000). Bowen (1999a) placed the Cromerian within OIS 21 to OIS 13, thus including four ‘glacial’ episodes (OIS 20, 18, 16 and 14) and four ‘interglacial episodes’ (OIS 21, 19, 17, 15), calibrated by amino acid and electron spin resonance (ESR) dating, whereas in Holland, three ‘glacial’ and four ‘interglacial’ episodes have been recognised for the same interval. True glacigenic sediments are absent in both successions, marine and fluviatile gravels, the latter derived from distant glacial sources, are interbedded with peats and organic muds. Both the Dutch and British sequences are rich in fossils (large and small mammals, mollusc shells, insects and plant remains) which indicate true warm and temperate interglacial conditions, as well as colder (interstadial) episodes in southern Britain and the Netherlands. Zagwijn (1985) gives estimated mean July temperatures of up to 20° C for the interglacial episodes and down to about 5° C for the interstadial ('glacial') episodes preserved in the Dutch sequence. Tills that have been assigned, by Bowen (1999b) to glaciation by a Welsh ice sheet in OIS 16, based on amino acid dating of fresh water molluscs in overlying fluvial sediments to OIS 15, are present south west of Bristol (Andrews et al., 1984), but this interpretation has been questioned subsequently (Scourse et al., 2009). Similar issues have arisen regarding the early 'glaciation of the upper Thames' and glaciation of the Scilly Isles. The latter has apparently been resolved by cosmogenic dating of morainic deposits associated with the glacial limit, which produced a post-Late Glacial Maximum (Post LGM) exposure age of 20.9 ± 2.2 to 22.1 ± 2.8 kyr BP consistent with deposition by the last (OIS 2) ice sheet (McCarroll et al., 2010); a view which is now generally accepted.

The Anglian Glaciation

The most widely accepted, early large scale glaciation of Britain is attributed to the (OIS 12) Anglian ice sheet (see Figure 21), which laid down extensive spreads of glacial till across East Anglia and the English Midlands and infilled buried valleys cut in bedrock with glacial deposits that locally exceed 100 m in thickness. The south eastern limit of the Anglian glaciation is well established in the Thames Basin, where the ice diverted the course of the early River Thames and in places blocked its drainage to form large proglacial lakes (Sumbler et al., 1996). The Anglian glacial limit in the Midlands corresponds with the southern limit of the glacial sediments within the Wolston Formation in Leicestershire, Warwickshire and adjacent areas. As their name implies, these deposits were originally assigned to the OIS 6 Wolstonian glaciations (Mitchell et al., 1973), but many have been reassigned to OIS 12 because they are overlain, in places, by Hoxnian (OIS 11) interglacial sediments (Maddy, 1999).

The stratigraphical correlation of the sequences originally assigned to the Wolstonian is, in fact, probably more complicated than this, as the sequences include glacial units that have now been tentatively attributed to OIS 6 and OIS 10 and terraced fluvial aggradations that have been assigned to many of the intervening warmer isotope stages, as well as to the OIS 8 cold stage. This situation is discussed at length by Catt *et. al.*(2006), who adopt an approach that seems appropriate to the present state of knowledge about this Quaternary interval in Britain. They ascribe the Wolstonian to a cold stage between the Hoxnian (OIS 11) and Ipswichian (OIS 5e) interglacials, and regard most of the fluvial aggradations as parts of a 'Wolstonian complex' that can be distinguished from the Hoxnian and Ipswichian sequences, in terms of their fossil floras and faunas, as well as by amino acid and ESR dating methods. The interval covered by the complex includes periods of significant climatic oscillation, spanning some 230,000 years, when two extensive glaciations occurred in northern Europe. Catt *et. al.*(2006) also describe the common situation, in central England and elsewhere, where poorly-dated glacial deposits have been attributed to the Wolstonian time interval, simply because they appear to be either older than the Ipswichian or younger than Hoxnian, or are lithologically distinct from nearby glacial sediments that have been attributed to OIS 12 or OIS 2 glacial events. Some may be overlain by Ipswichian or underlain by Hoxnian interglacial deposits, but they are never found between both. The acknowledged OIS 12 tills of East Anglia, for example, are overlain by interglacial deposits assigned to OIS 11 and OIS 5e and also by deposits, assigned to OIS 7, containing woolly mammoth, horse and aurochs. Most of the pre Holocene, interglacial sites in Cheshire appear to be of OIS 5e age, but this does not preclude underlying tills being formed during any pre Ipswichian cold stage.

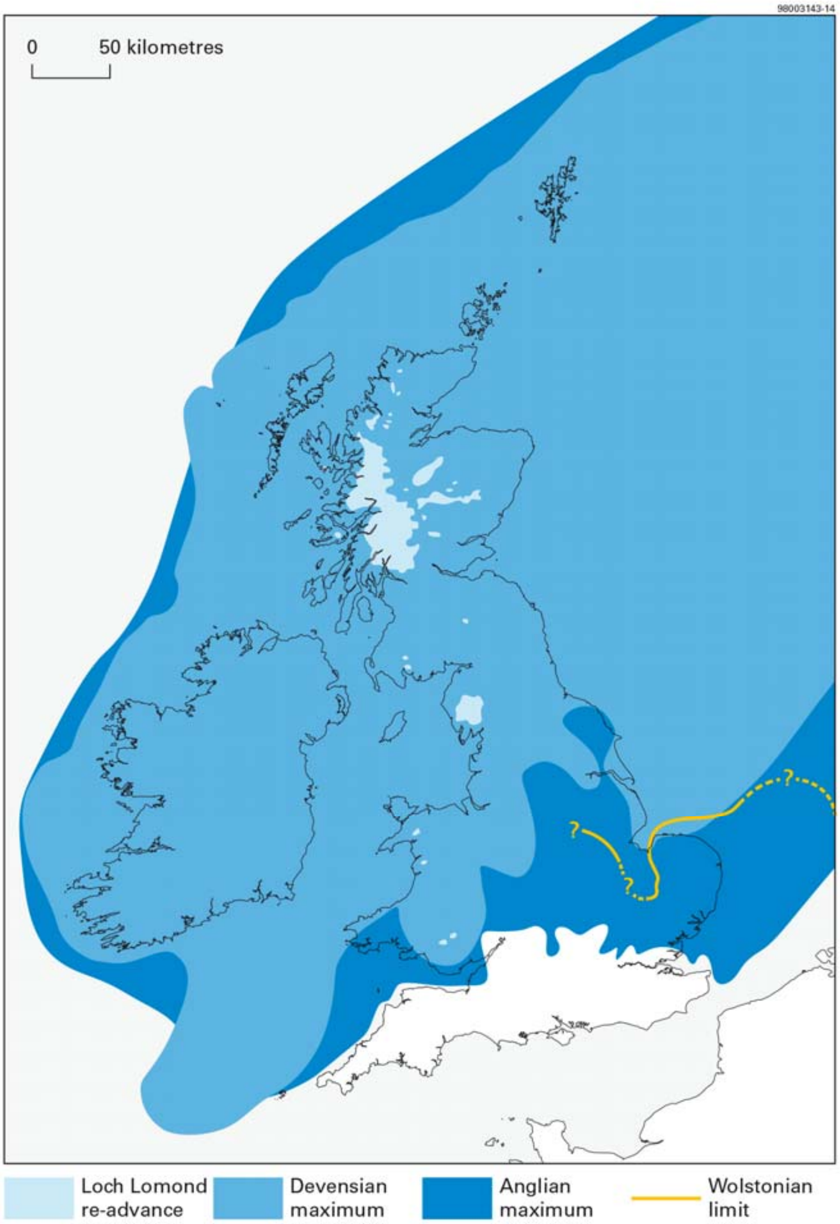


Figure 21. A reconstruction illustrating the maximum extent of ice cover during the Anglian, Main Late Devensian and Loch Lomond Stadial glaciations of the British Isles and the possible limit of Wolstonian glaciation in eastern England and the North Sea. (Modified from NDA, 2010a, prinicipally with data from reconstructions by Clark et. al, 2012a and Gibbard and Clark, 2011).

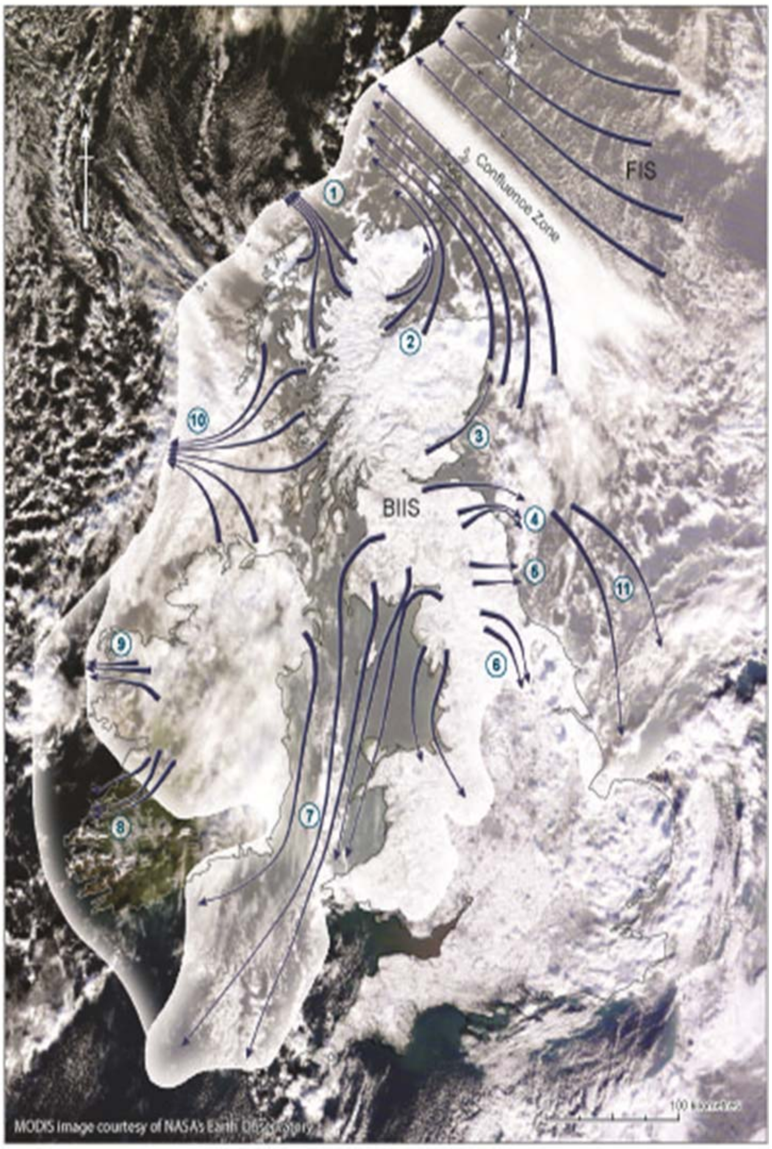


Figure 22. The maximum extent of the British and Irish Ice Sheet during OIS 2 and the positions of fast-flowing ice streams**.** NASA satellite image of the British Isles covered in snow and ice during the winter of 2009-2010. Shown on the image is the currently agreed maximum extent attained by the British and Irish Ice Sheet (BIIS) during the last ice age. The BIIS was drained at different times throughout the Late Devensian by several major ice streams: 1-Minch, 2-Moray Firth, 3- Strathmore, 4-Forth and Tweed, 5-Tyne Gap, 6-Vale of York, 7-Irish Sea, 8-Clare, 9-Clew Bay, 10-Sea of Hebrides, 11- North Sea. The largest of these was the Irish Sea ice stream which flowed from central Scotland as far south as the Isles of Scilly. To the north the BIIS meets with the Fennoscandian Ice Sheet (FIS). NASA image by Jeff Schmaltz, MODIS Rapid Response Team, Goddard Space Flight Center. Caption by Michon Scott.
http://eoimages.gsfc.nasa.gov/images/imagerecords/42000/42237/gbritain_tmo_2010007_lrg.jpg

The onshore extent of the Anglian glaciation north of its well defined limit in southern England is unclear. Fragmentary evidence of glacial sediments assigned to OIS 12 is present in Wales, but no clear western limit is recognised. It is probable that the limit lies offshore around the British Isles and that at its maximum extent grounded ice extended to the edge of the continental shelf, particularly to the north of Scotland, as it did during the OIS 2 glaciation (Figure 22).

Post Anglian -Pre Devensian sequences

In southern Britain, the Hoxnian interglacial warm period, that followed the decay of the Anglian ice, was generally characterised by temperatures that are slightly higher than those that occur at the present day and by higher sea-levels (> 20 m above those of the present day); some colder intervals are also normally thought to have been present. Most Hoxnian sites occur as fine-grained lake sediments that were formed in kettleholes or other depressions in the land surface. Some sites (including Purfleet in Essex and at Hoxne (after which this period is named) in Suffolk) have been assigned to OIS 9; though the OIS 9 age applied to the latter, from amino acid dating, is controversial. Other sites, notably Swanscombe in Kent have been assigned to OIS 11 (Bowen, 1999a). The most complete sequence has been recorded at Marks Tey in Essex (Turner, 1970), which has also been dated to OIS 11 (Rowe et al., 1999). Certain floristic characteristics, including a high frequency of buckthorn pollen, serve to distinguish Hoxnian pollen assemblages from those of other interglacial sites and, taken together, the flora indicates a more oceanic climate than at any other Quaternary interglacial stage. Although there is archaeological evidence, in the form of hand axes, of considerable human activity during and prior to the Anglian glaciation the Hoxnian represents the first time that there is widespread *in situ* evidence of man in the landscape of Britain (Wymer, 1988).

As mentioned above, many of the glacial deposits originally assigned to the OIS 6 Wolstonian glaciations (Mitchell et al., 1973), especially those in the English Midlands, have been reassigned to OIS 12 (Maddy, 1999). However, recent studies of glacial sequences on the Fenland margins between Peterborough and King’s Lynn (Gibbard et al., 2009) supported by luminescence dating of associated glaciofluvial deltaic deposits, indicates that an ice margin existed in that area at about 160 kyr (OIS 6). This would therefore represent part of the margin of a Wolstonian glaciation that the authors correlate across the North Sea with the equivalent OIS 6 Drenthe glacial limit in the Netherlands.

The next interglacial episode in Britain, for which extensive evidence is found, is the Ipswichian (OIS 5e), though isotopic dating and faunal differences indicate that there are some ‘Ipswichian’ sites that are OIS 7 in age. Only the early temperate part of the OIS 5e interglacial is preserved at the type site, at Bobbitshole near Ipswich, where the climate ranges of fossil beetles suggest summer temperatures were about 3°C higher than at present. The famous ‘Trafalgar Square Fauna’ from OIS 5e dated site in central London contains fossils of hippopotamus, lion, straight-tusked elephant, rhinoceros, fallow deer, red deer, giant deer and bison, suggesting a warm savannah climate; beetle faunas from similar sites suggest a Mediterranean climate. The ‘Ilfordian’ fauna from Ilford contains fossils of horse, mammoth, straight-tusked elephant and rhinoceros; a similar fauna is known from the OIS 7 site at Aveley in Essex. These suggest a generally cooler climate in OIS 7 than in OIS 5e, a conclusion supported by the absence of ‘Mediterranean’ beetles in the OIS 7 deposits.

Main Late Devensian Glaciation

In ‘classical’ British Quaternary stratigraphy, the Ipswichian interglacial is followed by the Main Late Devensian (OIS 2), which is commonly known as the last widespread glacial event that affected most of the UK landmass. This is now thought to have begun with the build-up of ice during the global LGM at c. 28 kyr BP (Clark et al., 2009). However, there is fragmentary, and often controversial evidence of older (pre-LGM) glacial episodes throughout parts of the Devensian Stage (OIS 4 – 2) (Bowen et al., 2002). Part of the difficulty is the concept of local use of the term LGM and applying it to the British Ice Sheet (BIS), and also to the British and Irish Ice Sheet (BIIS). Confusion also arises in the British case, by directly equating the LGM with the Dimlington Stadial/Dimlington Chronozone at 26 kyr ^{14}C BP – 13 kyr ^{14}C BP (Rose, 1985). In some instances however, there appears to be evidence of post Ipswichian glaciation prior to 30 kyr BP with cosmogenic dating of glacial erratics and glaciated rock surfaces in southwest Ireland indicating deglacial ages between 37 and 32 kyr BP (Bowen et al., 2002) and recently published paired isotope ages from erratics on Lundy in the Severn Estuary suggest that a deglacial episode affected the island between about 31 and 32 kyr BP (Rolfe et al., 2012).

Amino acid dating of marine molluscs in shelly tills has also been put forward to suggest that parts of north eastern Scotland (notably Caithness, Orkney and Buchan) were glaciated between OIS 5e and c. 37.5 kyr BP, but had remained ice free since that time (Bowen et al., 2002). Much of the evidence, which was based on ages from shells transported as glacial erratics within younger tills, for ice free areas in NE Scotland has since been refuted (Merritt et al., 2003; Phillips et al., 2008; Hall et al., 2011). Nevertheless, the recognition that the BIIS and the Fennoscandian ice sheets may have been confluent during the global LGM means that there may have been glaciation of Britain that commenced during the Devensian, before the beginning of the OIS 2 Dimlington Stadial (Sejrup et al., 2000). Although few undoubted glacial sediments can be attributed directly to Early or Middle Devensian glaciations (OIS 5d and 5c, OIS 4 and OIS 3), U- Series dating of organic interstadial sediments and luminescence dating of fluvio-aeaolian sands from Chelford in Cheshire (Heijnis and van der Plicht, 1992; Rendell et al., 1991), shows that these deposits were laid down in interstadials during the Early Devensian. Similar organic sediments, dated to c. 106 kyr BP occur between Late Devensian till and above an earlier till at Moy near Inverness (Heijnis and van der Plicht, 1992; Walker et al., 1992). This indicates that even in areas of the Scottish Highlands, where erosion by the last British ice sheet was extensive, remnants of the pre-Late Devensian glacial and interstadial sequences can be preserved.

The maximum extent of the BIIS during the Dimlington Stadial is now becoming quite well established in much of southern Britain. The general position of its onshore southern limit from Lincolnshire, through the English Midlands and the Welsh Marches is clearly established on geomorphological, lithological and stratigraphical grounds and the location of this boundary has changed little in modern reconstructions, since the pioneering work of Charlesworth (1957). The position of the limit in South Wales and the Irish Sea basin has, until recently, been the subject of much more scientific debate, based in part on the absence of adequate dating of possible positions. The advent of cosmogenic dates for ice limits in Pembrokeshire and the Scilly Isles has addressed this issue (McCarroll et al., 2010) and shows that the last BIIS ice sheet reached the northern coast of Scilly and probably extended further southward into the Celtic sea. The generally agreed ice limit position is shown on Figure 22. Some controversy remains however about its extent in south western Ireland, possibly as a consequence of the pre Late Devensian deglacial cosmogenic ages from this area mentioned previously. Recent detailed images of seabed bathymetry (Bradwell et al., 2008) suggest that the northern sector of the BIS reached the edge of the continental shelf, and most authors agree (Bowen et al., 1986; Bowen, 1999b; McCarroll et al., 2010; Bowen et al., 2002; Sejrup et al., 2005) that ice reached the shelf edge to the west of Ireland; they disagree about the age of that glacial event.

There is general agreement that a strong global cooling event began after about 30 kyr BP and culminated about 20 kyr BP. Ice expanded from the Scandinavian mountains to cover much of northern Europe to reach the vicinity of Copenhagen, Berlin and Moscow; permafrost developed to depths exceeding 100 m and extended as far as southern France (Boulton et al, 1991) and wind-blown loess blanketed parts of southern England. In Britain ice accumulation began in the Scottish Highlands and merged with other centres in the Lake District and Snowdonia and the resultant ice sheet is now thought to have been confluent with ice from Ireland and Scandinavia. There is controversy about when, and for how long the BIIS was confluent with the FIS and also regarding the pattern and timing of ice retreat during deglaciation (Rose et al., 2001; Clark et al., 2012a). This is important, because until very recently most numerical models of ice sheet dynamics and of ice thickness attributed to the last BIS, used a minimalist 'traditional' view of a single separate British ice sheet configuration, extent and deglaciation pattern which impacts on reconstructions of past ice loading, isostatic depression and rebound used to inform modelled future FEP scenarios for these drivers of environmental change in Britain.

The most up-to-date reconstructions (Clark et al., 2012b; Gibbard and Clark, 2011) suggest that a confluent ice sheet began to grow c. 30 kyr BP and reached its maximum extent prior to the onset of deglaciation at the shelf margin at c. 27 kyr BP. This is supported by cosmogenic dates from erratic boulders on South Rhona, close to the North-west continental shelf margin, which indicate that, at its maximum extent, Late Devensian ice over-rode the island and subsequently retreated at c. 25 kyr BP (Everest et al, 2012). Much of the North Sea basin was deglaciated by c. 25 kyr BP, but ice was still advancing at around this time in the Irish Sea basin, along the Lincolnshire coast and across the Midlands. Consequently, both glaciation and deglaciation of the British landmass are diachronous, with ice advancing in some areas and retreating in others, at the same time, which is why there is controversy as to when the BIS reached its maximum extent (traditionally c 20 kyr BP; the often quoted LGM for the BIS). The pattern of retreat is conventionally determined by the interpreted dating of geomorphological features (such as moraines) that mark ice limits (Boulton, 1992); this has now been enhanced by cosmogenic dating of erratics in the moraines themselves (Ballantyne, 2010). These limits delineate a complex pattern of retreat and ice sheet disintegration with ice receding back towards its original source areas. This retreat was punctuated by local ice readvances, particularly by fast flowing ice streams, such as that which occupied the Irish Sea basin (Figure 22). Most of southern Britain was deglaciated by about 18.5 kyr BP, though deglaciation of Scotland did not begin until about 15 kyr BP, when glaciomarine silts with arctic shelly fossil assemblages were deposited around the eastern coast. The timing of ice accumulation and the speed and pattern of retreat, outlined above, are a topic of much active research and discussion, and although many detailed inconsistencies remain to be resolved (Clark et al., 2012a), the balance of evidence indicates rapid expansion of ice to the edge of the continental shelf prior to the ' traditional' LGM at c. 20 kyr BP (Figure 23), followed by organized retreat commencing around 27 kyr and ending in most of Scotland by about 12.8 kyr BP,when marine silts containing high-boreal shell assemblages began to be laid down around the Clyde Estuary.

The Lateglacial Interstadial and the Loch Lomond Stadial

The period covering the decay of glaciers of the Main Late Devensian Glaciation and prior to the final growth of glaciers in Britain during the Loch Lomond Stadial is known as the Lateglacial (or Windermere) Interstadial (c. 15 kyr – 12.8 kyr BP). It coincides with rapid warming of the north eastern Atlantic, recorded in the GRIP ice core record from Greenland (Johnsen el al, 2001) and in sea bed sediment cores from the Hebridean shelf (Kroon et al., 1997). This warming commenced at about 15.5 kyr BP and ended about 14.3 kyr BP; it was followed by cooling that culminated in the Loch Lomond Stadial between 12.5 and 11.5 kyr

BP. These periods coincide with Greenland Interstadial 1 [GIS1] and Greenland Stadial 1 [GS1] from the GRIP record. During the interstadial, the atmospheric temperatures in Britain, as inferred from fossil beetle assemblages, rose abruptly to come close to (and in some instances exceed) those of the present day (Atkinson et al., 1987). The vegetation responded more slowly than the insects, to change from pioneer communities of heath and birch developed on the recently deglaciated land surface to become full temperate woodland in the southern Britain and in the valleys of Scotland; this change in vegetation was diachronous, with temperate plant communities migrating northwards as the temperature rose. Although little vegetational change appears to have occurred during the main part of the interstadial, beetle faunas indicate that temperatures were 2- 3°C colder than in its earliest phase.

The cooling that accompanied the re-growth of glaciers in the western Highlands of Scotland, during the Loch Lomond Stadial, also led to a major expansion of the FIS during the equivalent Younger Dryas Stadial. Both were associated with Heinrich Event H0. At this time, small valley glaciers and corrie glaciers formed in the uplands of western Britain (notably in the Lake District and Snowdonia). Permafrost was widespread in the unglaciated areas (Ballantyne and Harris, 1994) and wind-blown loess deposits were widely developed in Lincolnshire, the Vale of York and parts of Lancashire (Catt et al., 2006). In Scotland ice accumulation began in the Rannoch Moor area and the ice sheet expanded to cover the whole of the Western Highlands and smaller ice caps developed in Assynt, in the Grampian Highlands and on Skye, Mull and Arran (Golledge et al., 2008); recent research indicates that a plateau ice field and corrie glaciers also developed in the Mondaliath Mountains, south of Inverness (Gheorghiu et al., 2012; Boston, 2012). Many of the glacial limits have been bracketed by ^{14}C dates, taken at the base of peat sequences from beyond and within the bounding moraines; others have been identified by floristic changes in pollen spectra and in some more recent studies by cosmogenic dating of boulders within the moraines themselves. In the type area, around Loch Lomond, the limit is marked in many places by moraines which have incorporated marine shells from the high-boreal assemblages that were deposited in the Clyde Basin at the end of the Lateglacial Interstadial (Phillips et al., 2003).

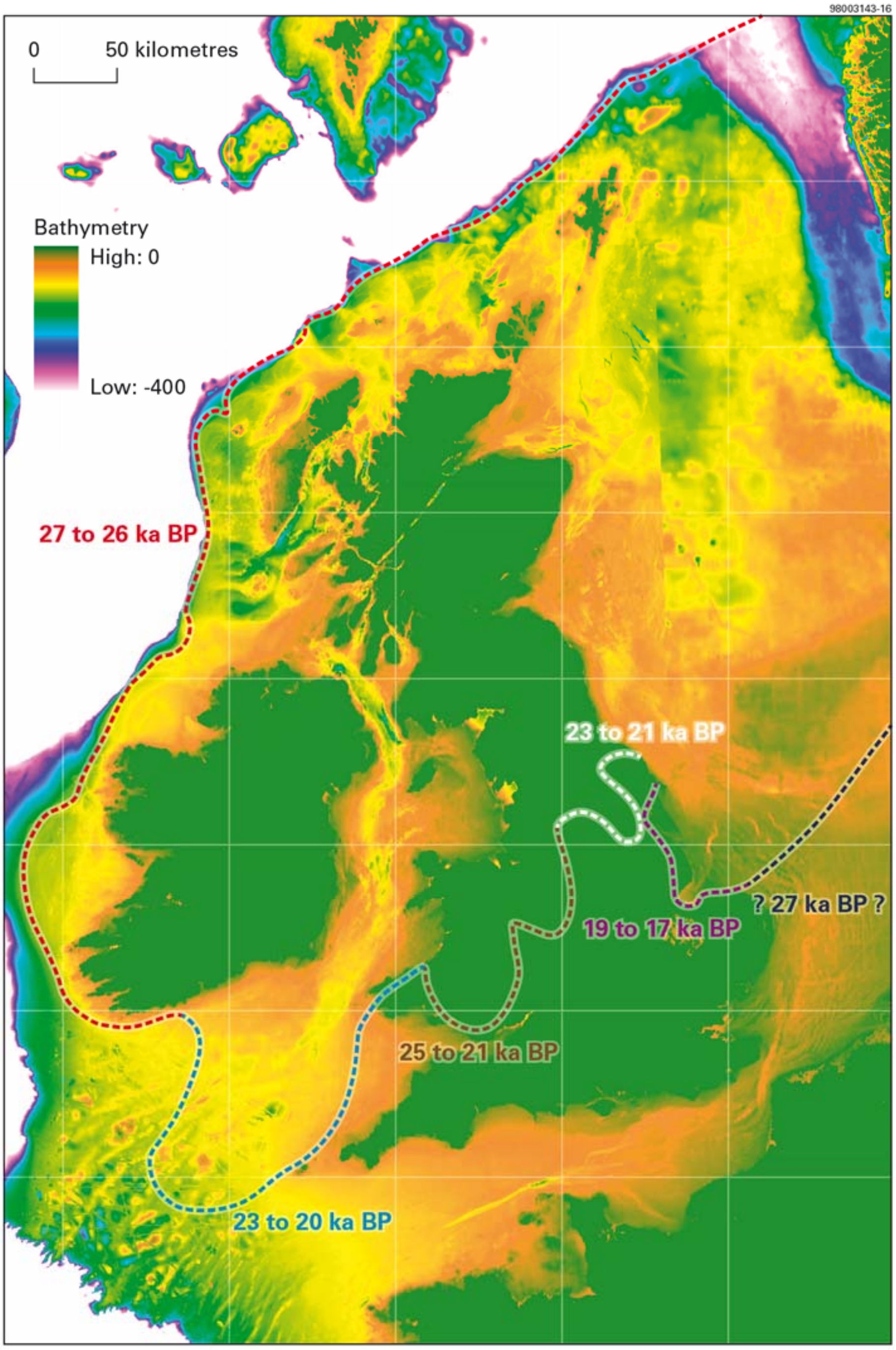


Figure 23. Latest reconstruction of dated confluent BIIS-FIS maximum limits, modified after Clark et al, (2012b).

The Holocene Interglacial

The climatic amelioration at the end of the Loch Lomond Stadial was abrupt and the rise in temperature may have been as great as 1°C per decade (Atkinson et al., 1987) at the beginning of the present Holocene (Flandrian) Interglacial (c. 11.5 kyr BP) which was comparable to the increase of c. 7°C in less than 50 years reported from Greenland ice cores. In Northern Europe, the Holocene period has been subdivided into five phases, based on pollen assemblages from lake sediments and peat bog sequences. These phases, which reflect a combination of climatic changes and human impacts are (in order of decreasing age): Pre-Boreal, Boreal, Atlantic, Sub Boreal and Sub Atlantic (vegetation zones IV-VIII of Godwin (1975)). Approximate time ranges for these divisions are: Pre- Boreal and Boreal ~ c. 11, 350 – 8,850 BP; the Atlantic and Sub Boreal ~ 8,850 – 2,587 BP; Sub Atlantic ~ post 2587 BP (Catt et al., 2006; Jones and Keen, 1993). Three chronozones (or substages) have also been proposed: a pre-temperate substage (F1); an early temperate substage (F2) and a late temperate substage (F3). The diachronous nature of the climate/vegetational zonation across the British Isles and Europe, and the plethora of terms applied to the subdivision of the Holocene significantly hinder coherent synthesis of proxy environmental records that lack precise radiometric (^{14}C dating) control. This diachronaity is well illustrated by high resolution proxy summer air temperature records derived from fossil insect assemblages, for the Lateglacial-Holocene transition in Scotland, which differ in detail from the trends observed in oxygen isotope records from Greenland ice cores, and from similar proxies in Scandinavia and other parts of northern and central Europe (Bell and Walker, 2005).

During the period of abrupt warming at the beginning of the Holocene, a rapid colonization of the deglaciated and periglacial lowland landscape by coniferous and subsequently by deciduous woodland took place. This continued until the mid-Holocene, when a possible combination of climate change, parasitic disease, and woodland clearance by man, led to a decline in several tree species. After this time, there is evidence of increased woodland clearance for agriculture from c. 6 kyr BP, after c. 5.1 kyr BP, again around 4.5 kyr BP and after c. 2.7 kyr BP (Catt et al., 2006). Several fossil proxy climate indicators (pollen, insects and fresh water molluscs) suggest that mean temperatures in England and Wales were 1- 3°C above present levels during the early and mid-Holocene (Bell and Walker, 2005). However, temperature deceased progressively after c. 5.7 kyr BP, culminating in the Little Ice Age of the 14th-18th centuries, when glaciers readvanced in the Alps and in Iceland, and lowland temperatures were 1-2°C lower than at present in Britain; with mountainous areas colder still.

Spectral analyses of bog surface wetness records in Britain, suggest cyclic periods (with a periodicity of c. 600, 800 and 1,100 ^{14}C years) of increased precipitation, beginning at c. 8.5 kyr BP and, in some cases, lasting several hundred years (Barber et al., 2003). These episodes of increased precipitation events have been linked to major river flooding events across Britain since c. 10.4 kyr BP (Macklin and Lewin, 2003). Episodes of increased aridity during the early and mid Holocene have been identified in stable isotope profiles from lake sediments in Greenland (Anderson and Leng, 2004), and from tufa records in Britain (Garnett et al., 2004), as well as from a variety of proxies in Spain and across the Mediterranean basin.

The most controversial signal of the anthropogenic influence on the climate of the Northern Hemisphere for the last millennium is the reconstructed temperature curve produced by Mann *et. al.* (1999) following an initial reconstruction of global temperature for the last six centuries (Mann et al., 1998) based on a variety of proxies and instrumental measurements. This 'hockey stick' graph, which has been adopted by many climate scientists, distinguishes the Little Ice Age and the Medieval Warm Period (c. 950 – 1,250 AD) and indicates a rapid warming at the beginning of the 20th century. Much discussion has ensued about its accuracy and validity (New Scientist, 2009; Mann, 2012; Montford, 2010), but its general thesis has been supported by subsequent climate reconstructions covering the last 2,000 years (Solomon et al., 2007).

4.4 WHAT WILL BE THE EFFECT OF (ANTHROPOGENIC) CLIMATE CHANGE ON TOP OF LONG RANGE FUTURE CLIMATE SCENARIOS?

Most early long-term future climate modelling suggested that, without the impact of anthropogenic effects, the 100 kyr Milankovich cyclicity that has dominated the past almost 800 ka, will continue. It also indicated that global cooling would take place for the next 23 kyr (Imbrie and Imbrie, 1980) and glaciation of the land masses in the Northern Hemisphere would occur within the next 100 ka. Raymo's (1997) global climate curve (Figure 19), for example, predicts that the present Holocene interglacial is nearly over and that an intense "double peaked" glaciation would follow, terminating about 64 kyr AP (after the present); a recent study suggests that without CO_2 forcing that the current interglacial would end within the next 1,500 years (Tzedakis et al., 2012).

Because the variations in Earth's orbit that control the amount of solar energy (insolation) that reaches the surface can be predicted by Milankovich Theory (see Section 1.1), global insolation can be fairly precisely calculated for the next 130 kyr (Bretagnon and Francou, 1988). Calculations further into the future are not so reliable.

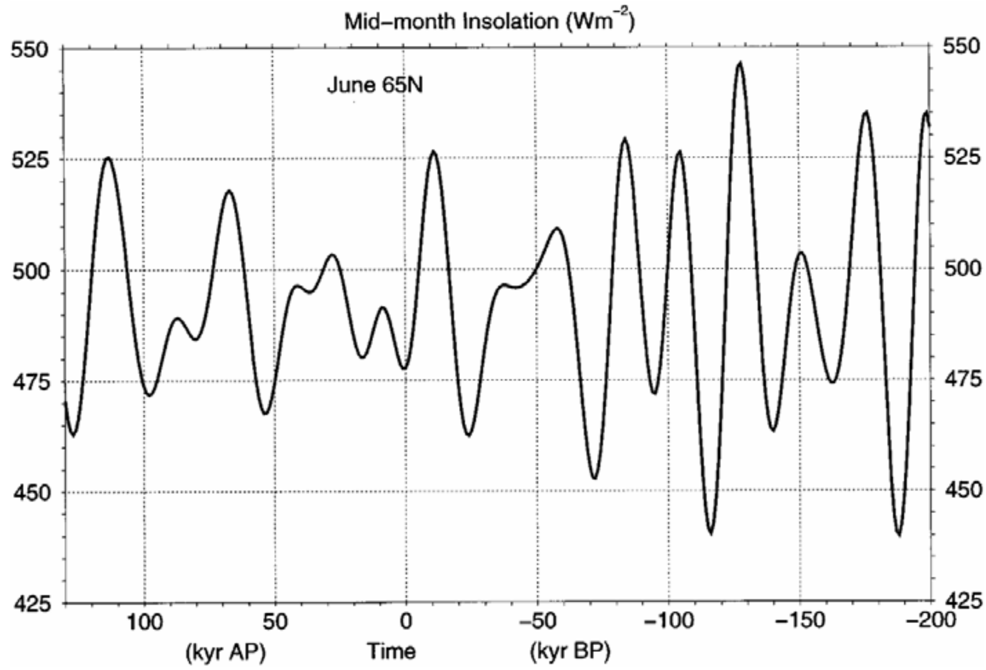


Figure 24. Model of June midmonth insolation at 65° N over the past 200 kyr and next 130 kyr from Loutre and Berger (2000).

The major feature of the calculated insolation for the next 130 kyr is the small amplitude of its variation (Figure 24). This insolation variation between 5 kyr BP and 60 kyr AP is exceptional, and has few past precedents (Berger and Loutre, 1996; Berger et al., 1996). Most models assume that insolation is the principal driving force for the climate system and some suggest the best and closest analogue for our near future climate is that of OIS 11 (the Hoxnian Interglacial; c. 360 - 420 kyr BP) (Imbrie et al., 1984; Berger and Loutre, 2003; Loutre and Berger, 2003). Others (Tzedakis et al., 2012) have argued that Marine Isotope sub-stage (MIsS = OIsS) 19c; an interglacial during the Cromerian, is a closer analogue. Whatever the detailed differences of interpretation of past analogues, most climatologists agree that high concentrations of atmospheric CO_2 will raise global temperatures and delay the onset of glaciations (Berger and Loutre, 2002).

4.4.1 The effect of greenhouse gases and aerosols

Greenhouse gases and aerosols in the atmosphere respectively increase and decrease the amount of solar radiation hitting the surface of the Earth. Both have increased due to human activity. Consequently, it is important to understand the effect of greenhouse gases and aerosols in predictions of future climate.

Figure 25 shows the radiative forcing of many external factors, including aerosols and greenhouse gases (CO_2 and CH_4) and the degree of scientific understanding of each of these factors. Except for solar variation some form of human activity is linked to each. The effect of CO_2 and other greenhouse gases, such as CH_4 and NO_2 is very well understood, whereas aerosols and their role in the climate system are relatively poorly understood.

It is widely believed that anthropogenic (human induced) CO_2 and CH_4 began to change the natural concentration of greenhouse gases as industrialisation advanced at the beginning of the 19th century. However, it has also been suggested that mankind has been influencing the climate for perhaps the last five thousand years (Ruddiman, 2003b; Ruddiman, 2005), and that, without human intervention, the Little Ice Age climate would have persisted until the present day. CO_2 and CH_4 emissions caused by early agriculture, are believed to have augmented global mean temperature by ~0.8 ºC, and the temperatures at higher latitudes by roughly 2.0 ºC. These rises in temperature would have been sufficient to have halted an oncoming glaciation.

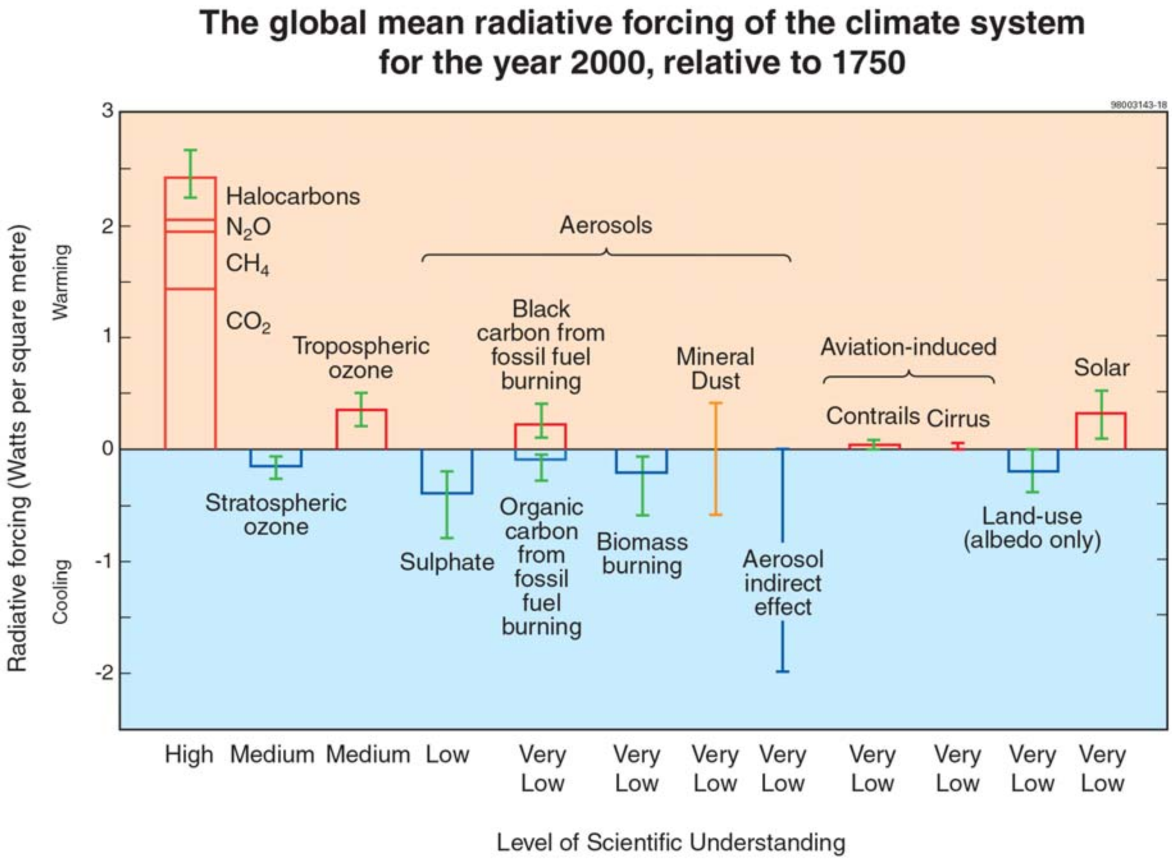


Figure 25. Radiative forcing of external factors. Vertical lines associated with boxes indicate a range of estimates, guided by the spread in the published values of the forcings, and by the degree of scientific understanding. A vertical line alone denotes a forcing for which no best estimate can be given; modified after Cedercreutz (2004).

The main trends in CO_2 and CH_4 concentrations are said to be similar for all glacial-interglacial cycles (Petit et al., 1999; Kukla et al., 2002). At glacial-interglacial transitions, concentrations of CO_2 rise from 180 to 280-300 parts per million (ppm) and that of CH_4 from 320-350 to 650-770 parts per billion (ppb). The decrease of CO_2 to the minimum value, during the onset of glaciation is slower than its increase towards interglacial values. The CO_2 decrease lags the temperature decrease by several thousand years. The greenhouse gas record shows that present-day levels of CO_2 and CH_4 (~360 ppm and ~1700 ppb, respectively) have not occurred for at least the past 420 ka. Pre-industrialisation levels, however, are inferred from all interglacials (Cedercreutz, 2004).

Aerosols are small liquid or solid particles suspended in the atmosphere. They include sulphates, nitrates, organics, soot and dust. Some occur naturally, originating from volcanoes, dust storms, forest and grassland fires, etc. About 10% of aerosols derive from human activities, such as the burning of fossil fuels. Their important characteristic is that they have short atmospheric lifetimes, typically just a few days, whereas greenhouse gases persist in the atmosphere for tens to thousands of years. Exceptions can occur, however, when aerosols such as volcanic ash and sulphur dioxide are injected into the atmosphere by volcanoes, they can affect global climate for many years. For example, recent research has linked volcanic activity to the initiation of the Little Ice Age (Miller et al., 2012).

Aerosols increase the reflection of solar radiation back to space through a variety of complex radiative processes. In contrast to greenhouse gases, which are distributed globally, aerosols are generally regionally concentrated near anthropogenic sources, mostly in the Northern Hemisphere. Because most aerosols reflect sunlight back into space, they tend to cause local cooling of the Earth's surface. It is suggested that this may partially offset the effects of global warming, in regions with high aerosol concentrations. Modelling indicates that global surface warming from greenhouse gases should exceed the global cooling effect of aerosols; regionally however, the cooling effect of aerosols could exceed the short-term effects of greenhouse-gas warming (Ramanathan et al., 2001).

4.5 MODELLING RESEARCH ON GLOBAL FUTURE CLIMATE CHANGES

Most early models of future climates infer the timing and duration of future glacial, interstadial and interglacial climate, based on their fit to data from the last 800 kyr and restrict this to considering possible global climates into the next 100to 130 ka; they deliberately discount anthropogenic forcing. These model projections of future climate change were based on simplified simulations of the past global climate. This was partly due to limited computing resources, but also to a rudimentary understanding of certain (nonlinear) feedback mechanisms within the climate system.

4.5.1 Simple Types of Long Term Global Models

These are typically Orbital Forcing Models, based on future projection of past trends of Milankovitch cyclicity. These typically predict global climate states 100 kyr to 1 Myr into the future. Examples of these types of models are those of Imbrie and Imbrie (1980) and the Astronomical Climate Index Model (ACLIN) models of Kukla *et. al.* (1981); all avoid consideration of feedbacks within the climate system and directly correlate orbital parameters to proxy data. The former uses radiometrically dated estimates of global ice sheet minima and maxima and OIS boundaries, and $\delta\ O^{18}$ curves from deep sea cores for the last 500 ka; the latter were calibrated by pollen records, sea-level data and $\delta\ O^{18}$ curves from the last 730 ka.

4.5.2 Intermediate Complexity Models (Coupled Ice Sheet Models)

These are typically coupled climate-ice sheet models, such as the LLN 2D NH climate model. This is a coupled sectorially averaged climate-ice sheet model, composed of a two-dimensional model of the Northern Hemisphere coupled to an ice sheet model which can be used to investigate the relative importance of components of the climate system in the transition between glacial and interglacial conditions (Berger et al., 1996; Gallée et al., 1991; Gallée et al., 1992). The model was specifically designed to simulate and investigate long-term climate variations in response to Milankovitch forcing. It links the atmosphere, the upper mixed layer of the ocean, sea ice and the continental, ice sheets of the Northern Hemisphere. It is forced by variations in insolation, precipitation, evaporation, vertical heat fluxes, surface albedo (capacity to reflect), oceanic heat transport and oceanic mixed layer dynamics, but does not include a carbon cycle, or CO_2 concentration in the atmosphere. Based on past ice sheet reconstructions, the LLN 2D NH predicts past melting of Northern Hemisphere ice sheets (notably the Greenland Ice Sheet) on too many occasions, but it does indicate that OIS 11 is a suitable analogue for future natural climate change (Loutre, 2000).

Various calculations have been made using the LLN 2D NH model to simulate climate for the next 130 ka, with representations of present-day Northern Hemisphere ice sheets and different scenarios for future CO_2 concentrations (Loutre and Berger, 2000). Most of the natural scenarios (no fossil fuel induced CO_2 contribution) indicate that the Earth's climate is likely to experience long-lasting, ~50 kyr interglacial conditions and that it will not enter naturally into the next glaciation before 50 kyr AP. Without anthropogenic forcing, the next interglacial maximum is expected to be most intense at c. 100 kyr AP, and an interstadial is likely at c. 60 kyr AP. After 100 kyr AP, continental ice will rapidly melt, leading to a minimum Northern Hemisphere ice volume at 120 kyr AP.

Over longer time scales, with a large fossil-fuel induced CO_2 value included, the LLN 2D NH model shows almost complete melting of the Northern Hemisphere continental ice cover during the next 150 ka. Beyond 500 kyr AP, the fossil fuel contribution to global warming becomes smaller and the climate would recover to its natural state. With an intermediate level of fossil-fuel induced CO_2 gradually declining (Figure 25), the model indicates, that the Greenland Ice Sheet will disappear in the next few thousand years and that the next episode of widespread glaciation of the Northern Hemisphere should not be expected before c. 170-180 kyr AP.

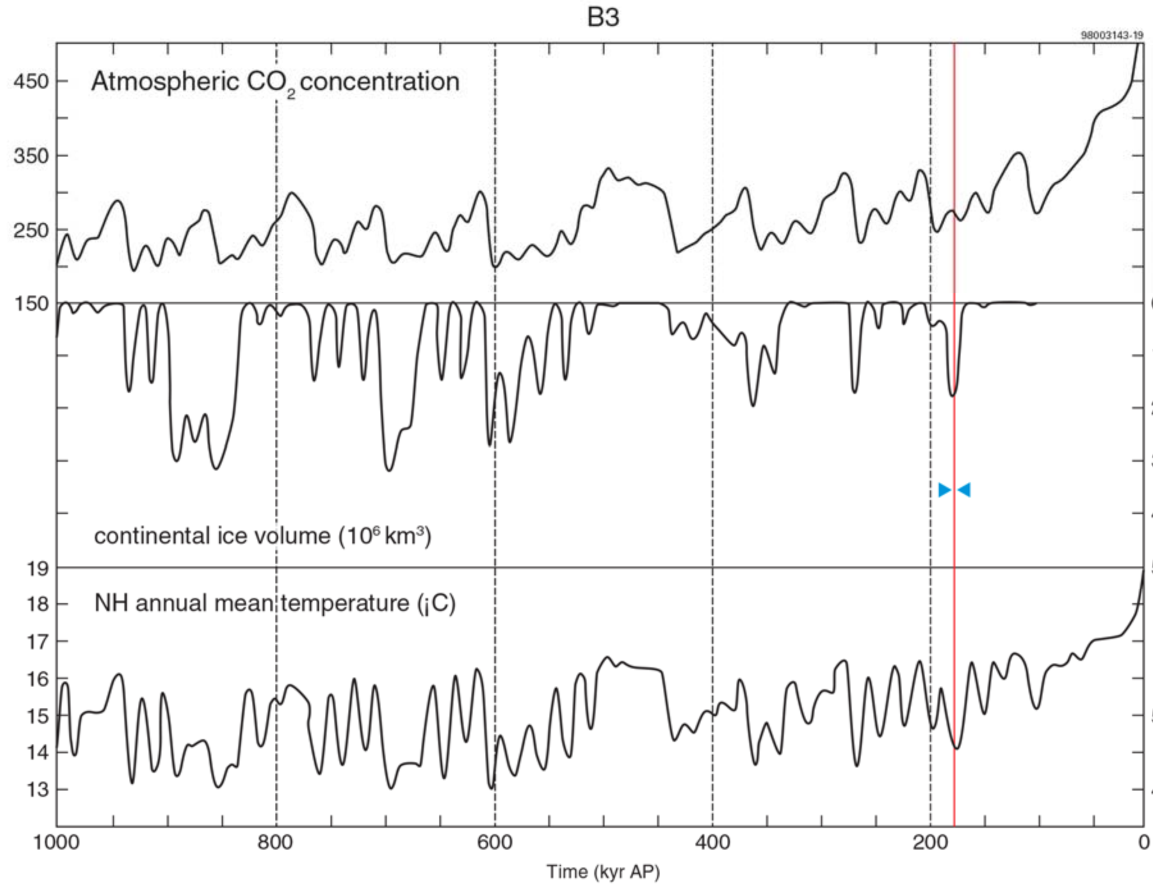


Figure 26. Northern Hemisphere future model to 1Myr with Intermediate CO_2 forcing. The red line shows the first major glaciation at c. 170-180 kyr AP (modified after BIOCLIM (2001).

Two other principal Earth Systems Models of Intermediate Complexity [ESMIC], MoBidiC and CLIMBER-GREMLINS, which link ice sheet and climate models, have been used as part of the BIOCLIM project. This modelled sequential BIOsphere systems under CLIMate change for radioactive waste disposal, as part of the EURATOM fifth European Framework programme. It aimed to provide a scientific basis and practical methodology for assessing the possible long term impacts on the safety of buried radioactive waste GDFs, with respect to climate and environmental change. The BIOCLIM results are presented in 12 deliverables from 5 work packages (BIOCLIM website). Of these, the scenarios for CO_2 concentrations, and insolation forcing during the next million years (BIOCLIM, 2001); continuous climate scenarios over western Europe (1000 km scale), (BIOCLIM, 2003a); regional climatic characteristics at specific times (BIOCLIM, 2003b); development rule-based down scaling for the MoBidiC model (BIOCLIM, 2003c) and development of the physical/statistical downscaling methods to the CLIMBER climate model (BIOCLIM, 2003d) are the most important for the modelling of future climate states applicable to the construction of a deep GDF in Britain.

The CO_2 emissions and atmospheric concentration scenarios, derived by the BIOCLIM project, were initially applied to the LLN2-D NH model (BIOCLIM, 2001) and subsequently to the MoBidiC, GREMLINS and CLIMBER-GREMLINS models (BIOCLIM, 2003a). In MoBidiC the ice sheet model used was rather simple, while it was more complex in CLIMBER and CLIMBER-GREMLINS. All of the models account for atmosphere-ocean-vegetation-ice sheet interactions and, because the vegetation model used is the same in all of the climate-ice models, its output can therefore be used to compare the effects of climate changes in each. Several simulations were performed using both CLIMBER and MoBidiC, to evaluate the ability of each to simulate past climates. Snapshot simulations were performed to allow comparison with General Circulation Model experiments, in particular with the results from the Palaeoclimate Modelling Intercomparison Project (PMIP); see below. The experiments were: A snapshot experiment of present-day climate; snapshots of extreme climate during the last glacial-interglacial cycle, i.e. the Last Interglacial (taken as 126 kyr BP); Last Glacial Maximum (taken as 21 kyr BP) and the Holocene Climate Optimum (taken as 6 kyr BP). Transient experiments were taken for the Last Interglacial (126 kyr BP-115 kyr BP); the Last Deglaciation (21kyr BP - 0) and the Last Glacial Cycle (126 kyr BP - 0). This last experiment was used to set the rules for downscaling complexity to the regional scale (BIOCLIM, 2003c).

Several global simulations, covering the next 200 ka, were performed using CLIMBER-GREMLINS and MoBidiC simulations and downscaled to provide regional climate estimates for different areas. Three scenarios (Figure 26) were presented for future variations in concentrations of atmospheric carbon dioxide (taking the contribution from fossil fuel combustion, combined with the projected natural variations) (NDA, 2010b). They are:

- Climate Scenario 1: Natural variations only with no further contribution from fossil fuel combustion;
- Climate Scenario 2: Natural variations plus a contribution from the low future use of fossil fuels; and
- Climate Scenario 3: Natural variations plus a contribution from the high future use of fossil fuels.

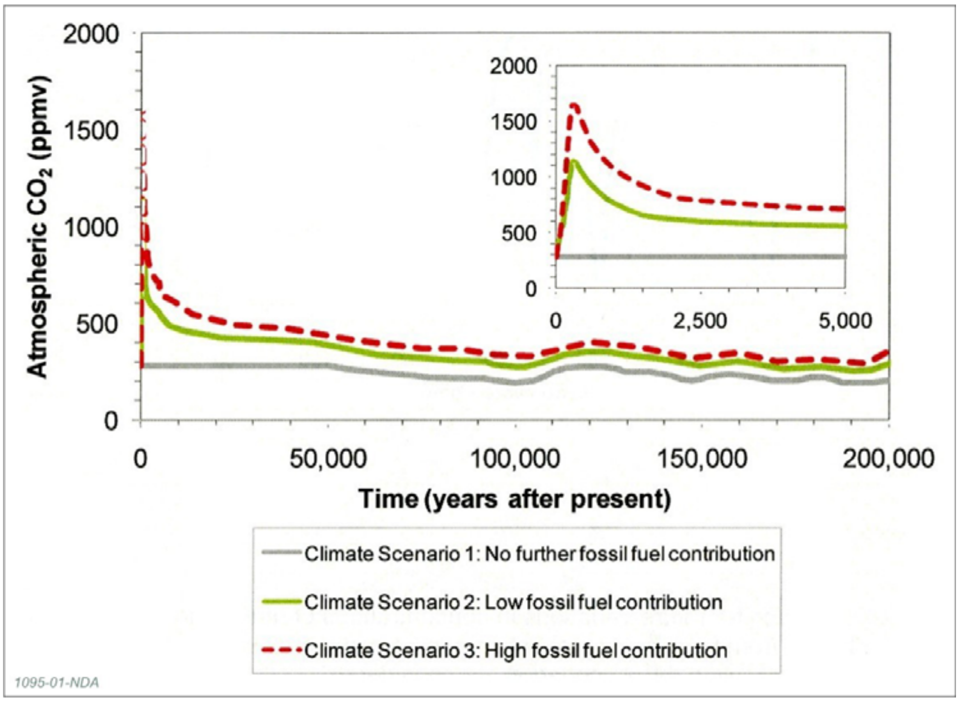


Figure 27. Projected global atmospheric CO_2 concentrations. Figure abstracted from BIOCLIM (2003a). The inserted chart provides increased resolution over the first 5 ka. The decrease in concentrations observed in Climate Scenarios 2 and 3 after a few hundred years occurs because fossil fuel reserves are largely exhausted or cease to be used.

The BIOCLIM modelling strategy involved defining sequences of climate states in specific regions, such as central England, for each of the climate scenarios studied (Texier et al., 2003).

Climate Scenario 1 (no fossil fuel contribution)

Figure **28** shows the projected climate states (summarized in Table 1) for central England, under the three scenarios[2]. Figure 29 shows the historical and projected northern hemisphere ice volume for Climate Scenario 1. These figures show that under natural conditions with no additional CO_2 forcing from fossil fuel combustion, that the next period of full glacial conditions is not projected to occur under this scenario for 100,000 years. The lower volume of northern hemisphere ice at that time in comparison with the last British on-shore glacial maximum, c. 18 – 20 kyr BP, strongly suggests that ice sheet formation in the British Isles would be limited to the north-western upland areas, as it was in the Loch Lomond Stadial and that lowland Britain would not be glaciated. This view is further supported by the shorter duration (< 10 ka) of the projected glacial episode compared with the Main Late Devensian glaciation (> 15 ka), as there would be less time available for British ice sheets to develop and spread south from their projected origins in the Scottish Highlands, the Lake District and North Wales. This again is similar to the situation during the Loch Lomond Stadial (which had a somewhat shorter duration, c. 1 ka). In lowland Britain, polar tundra conditions would be expected to prevail for a considerable time (c. 50 ka) prior to the onset of glaciations. The mean annual temperature of polar tundra conditions is consistent with the development of discontinuous (patchy) permafrost in central England. This is in accord with evidence from the past Quaternary glaciations, as relict permafrost features are widely preserved beyond the limits of previous glacial episodes (Ballantyne and Harris, 1994). Indeed, it is likely that discontinuous permafrost would begin to develop during the latter part of the preceding sub-arctic climate state (Harris, 2002), but it is also probable that short lived warmer periods, typical of most Quaternary interstadial episodes would also occur at this time, but these are below the resolution of the model.

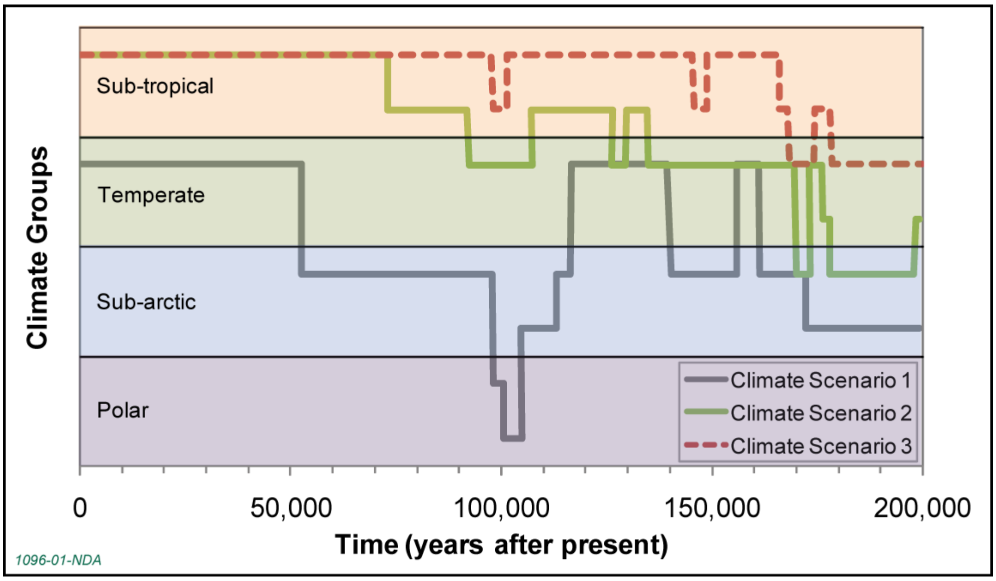


Figure 28. Projected future climate states for central England under the different scenarios. Figure based on BIOCLIM (2003c). Climate scenario 1: No further fossil fuel contribution. Climate Scenario 2: Low fossil fuel contribution. Climate scenario 3: High fossil fuel contribution

[2] Note that these sequences are derived from the output of the MoBidiC model used in BIOCLIM.

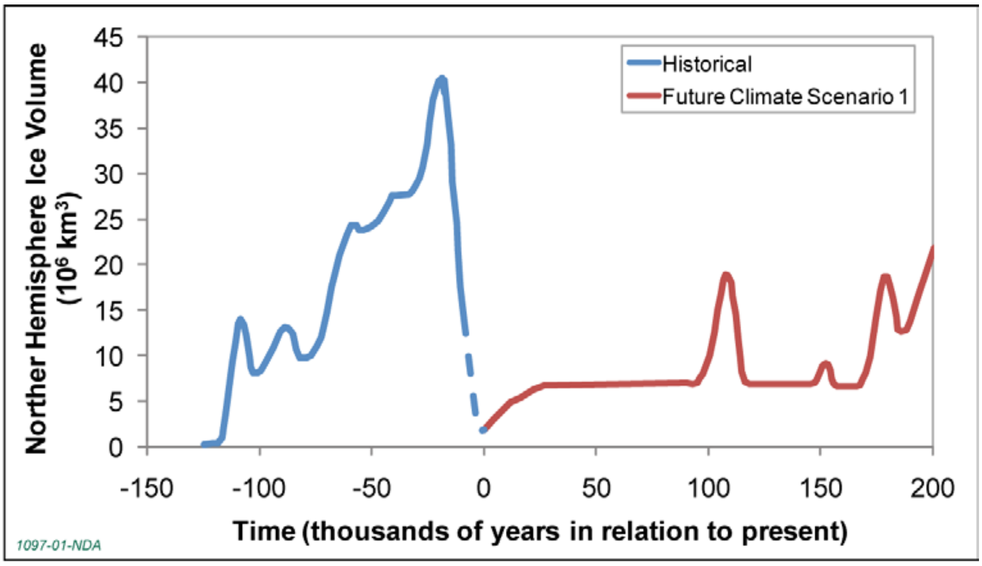


Figure 29. Modelled historical and projected northern hemisphere ice volume for Climate Scenario 1: No further fossil fuel contribution. Figure based on BIOCLIM (2003a).

Warming from polar tundra conditions is projected to occur somewhat more slowly than at the end of the Main Late Devensian glaciation. In that episode, the British ice sheet was at its maximum onshore extent between about 18 – 20 kyr BP and fully interglacial conditions were established in central England by about 11 kyr BP, during the early Holocene. However within the Late glacial period, substantial oscillations in climate occurred, notably the intensely cold but brief Younger Dryas/Loch Lomond Stadial, when upland glaciation occurred. In the projection, the warming to full interglacial conditions is indicated as occurring over a period of about 12,000 years. However, this distinction in timing is almost beyond the limit of the resolution of the models used.

Climate Scenarios 2 and 3 (low and high fossil fuel contributions)

In Scenarios 2 and 3 the effects of greenhouse-gas releases from fossil fuel combustion over the next few hundred years, see Figure 27, are expected to result in a substantial change in the British climate. The location of the British Isles, on the margin between oceanic and continental climate domains and the results of the different downscaling and modelling techniques adopted, imply a relatively rapid transition to cool sub-tropical conditions, see Figure 27. The results suggest that overall annual temperatures would be increased by about 2 to 3ºC, without any substantial change in the seasonal cycle and that the total annual precipitation would be very similar to that at the present-day (600 mm (Hulme, 2001)), but with a decrease of about 40 mm during the summer months. This appears comparable to the climatic conditions inferred for the Hoxnian (OIS 11) Interglacial in Southern Britain, and implies that the hydrologically effective rainfall (precipitation minus actual evapotranspiration) would be somewhat reduced relative to the present day, so increasing irrigation requirements.

Overall, following a peak in mean annual temperature over the next few hundred years, it is considered that a cooling trend would ensue in both Scenarios 2 and 3. This would result in temperate conditions similar to those of the present-day recurring between 90 kyr AP and 170 kyr AP. Thereafter, the modelling showed no strong trend in climate up to 200 kyr AP, although results for Scenario 2 indicated that a brief cold episode would be expected to occur at around 175 kyr AP and persist for a few thousand years. Beyond 200 kyr AP, the effects of greenhouse warming due to combustion of fossil fuels would decline to negligible levels. The BIOCLIM simulations for the Northern Hemisphere covering the period from 200 kyr AP to 1 Myr AP indicate that 'normal' glacial-interglacial cycling would return at 100 kyr periodicities and persist throughout that time interval.

Subsequent studies

Because the LLN 2D NH, MoBidiC and CLIMBER-GREMLINS models allow multi-millennial, scaleable simulations of future climate, that can be compared with past proxy climate records, they have been widely used in climatic impact assessments for the development of radioactive waste facilities in Finland (Cedercreutz, 2004), Belgium (Van Geet et al., 2009) and in the Meuse/Haute-Marne region of France (Andrieu-Ponel et al., 2007). They have also been used to predict the potential evolution of the climate and landscape of the Toledo area of Spain for the next 200 kyr (Recreo et al., 2005) and to inform ice sheet modelling, GIA modelling and permafrost modelling, associated with the post-closure safety case for a radioactive GDF in Sweden (SKB, 2006b).

Further research on the impacts of modelled climate changes in central England, for the next 200 ka under the three scenarios have been undertaken (Thorne and Kane, 2006; Thorne, 2010). These studies considered landscape development, including the impact on rates and processes of denudation, river incision and aggradation, and drainage basin evolution, as well as rates and modes of radionuclide transport, sea-level changes, lithospheric processes (principally isostatic adjustments), changes to vegetation and the impact on human communities.

What is evident, however, is that many of the climate narratives developed for central England are not directly applicable, without calibration, to much of upland Britain, particularly in the northern and western parts of the country. Here, particularly in the mountainous areas, the present day interglacial climate is more severe than in the lowlands, and more typical of some periglacial episodes that affected the lowlands during the Lateglacial Interstadial. This is indicated by the active development of periglacial features, such as patterned ground and solifluction lobes in Snowdownia and the Scottish Highlands today (Ballantyne, 1987).

4.5.3 Intermediate Complexity Coupled Ocean - Atmosphere – Vegetation Models

These models tend to reconstruct detailed global palaeoclimatic conditions for specific time periods, and as such they can provide baselines for scalable models, such as those used in the BIOCLIM studies. Perhaps the most widely used are the PMIP (Palaeoclimate Modelling Intercomparison Project) 1, 2 and 3 Models. These models tend to reconstruct detailed palaeoclimatic conditions for specific time periods in the past, e.g. Last Glacial Maximum (Pinot et al., 1999; Braconnot et al., 2007) and Holocene (Wanner et al., 2008). The first PMIP model was designed to test the atmospheric component of atmospheric general circulation models (AGCM's) for the LGM (in this case taken as c. 21 kyr BP) and the mid Holocene climate optimum (c. 6 kyr BP) (PMIP, 2000). PMIP 2 investigated the role of climate feedbacks arising from the interactions of atmosphere, ocean, land surface, sea ice and land ice, for the same two time intervals; PMIP3 has extended the time intervals to include modelling of the last 1,000 years. A similar approach, employing an extensive range of Earth System Models (ESMs) for simulation of ice age cycles and long-term anthropogenic global change, over multi-millennial timescales has resulted in the Grid Enabled Integrated Earth System suite of Model(s) (GENIE) which began in 2003. The outputs have included modelled past climate (Holden et al., 2010), future climate (Lenton, 2006), and linked models of future climate and socioeconomic change (Drouet et al., 2006).

4.5.4 Atmosphere – Ocean Circulation Models

These are the most complex, detailed and tested models with multiple input scenarios (used by IPCC) (Solomon et al., 2007). These are commonly general circulation models (GCM's) which are mathematical models of the general circulation of the atmosphere and/or ocean and based on Navier–Stokes equations on a rotating sphere with thermodynamic terms for various energy sources (radiation, latent heat). These complex models are widely applied to weather forecasting, understanding short term (decade to century time scale) climate, and projecting climate change. These computationally intensive numerical models are generally restricted to predictions of climate to 3000 AD. Much regional research linked to specific climatic or weather–related phenomena (e.g. changes to monsoon regime, El Niño, drought prediction etc.) uses these types of models (Meehl et al., 2007). The climate projections for Britain are included within the IPCC regional climate projections for Europe and the Mediterranean (Christensen et al., 2007), which inform the short term modelling of future British climate outlined below.

4.5.5 UK Climate Models

The national-scale research is described in the UKCP09 Report UK Climate Projections web site http://ukclimateprojections.defra.gov.uk/21678 which is frequently updated. It uses a version of the HadCM3 Global Climate Model that uses a slab ocean that represents only the top 50 metres of the ocean as one layer and prescribes the effects of ocean heat transport rather than simulating ocean currents explicitly. The atmospheric component of HadCM3 can be run coupled with a slab ocean, this is known as HadSM3. This is faster and less computationally intensive than running the fully-coupled model, however, it lacks the ocean feedbacks of a fully-coupled model. For most of its climate trends, UKCP09 uses 1961 to 1990 as its baseline period (Jenkins et al., 2008). This means that the changes are reported relative to the average climate experienced between 1961 and 1990. Observed climate information, based on instrumental records, provides a 25 km resolution gridded dataset which shows the baseline climate for the UK. It includes temperature and precipitation data for central England from 1772 to 1992 AD (extended to 2012) (Parker et al., 1992) and this is used to calibrate future climate change for the next several decades or perhaps hundreds of years under high, medium and low emissions scenarios. These correspond to the A1FI, A1B and B1 IPCC Special Report Emissions Scenarios (SRES) (Nakicenovic and Swart, 2000). The High and Low emission scenarios are the same as those of the same name used in the United Kingdom Climate Impacts Programme (UKCIP02) (Table 2) (UKCIP02, 2002).

Special Report Emissions Scenarios (SRES)	UKCIP02 Climate Change Scenario	Storyline	Increase in Global Temperature (ºC)	Atmospheric CO_2 Concentration (ppm)
B1	**Low Emissions**	Global population peak in 2050; rapid change toward a service and information economy; introduction of clean and resource-efficient technologies; Global solutions to economic, social and environmental sustainability.	2.0	525
B2	**Medium-Low Emissions**	Continuously increasing global population (lower than A2) intermediate levels of economic development; less rapid and more diverse technological change than in the A1F and B1; local solutions to economic, social and environmental solutions	2.3	562
A2	**Medium-High Emissions**	Continuously increasing population; slow, fragmented technological change; regionally oriented economic development and per capita economic growth; preservation of local identities and regional differences	3.3	715
A1FI	**High Emissions**	Global population peak in 2050; rapid introduction of new and more efficient technologies but with fossil fuel intensive economy; convergence of per capita income between regions	3.9	810

Table 3. Changes in global temperature (ºC) and atmospheric carbon dioxide concentration (ppm) for the period (2071-2100 average) for the four scenarios used in UKCIP02 (2002) (CO_2 concentration in 2001 was about 370 ppm)**.**

They span almost the full range of SRES. SRES A2 and B2 'storylines' (grey highlight in Table 2), with higher, continuously increasing population scenarios were not used in UKCP09, as the population assumed is significantly higher than the high end of current projections. The UKCIP climate of the UK and recent trends report (Jenkins et al., 2008) notes that:

- Global average temperatures having risen by nearly 0.8 ºC since the late 19th century, and have been rising at about 0.2 ºC/decade over the past 25 years;
- Global sea-level rise has accelerated between mid-19th century and mid-20th century, and is now about 3mm per year;
- Central England temperature has risen by c. 1.0 ºC since the 1970s; temperatures in Scotland and Northern Ireland have risen by c. 0.8 ºC since about 1980;
- Annual mean precipitation over England and Wales has not changed significantly since records began in 1766. Seasonal rainfall is highly variable, but appears to have decreased in summer and increased in winter, although with little change in the latter over the last 50 years. All regions of the UK have experienced an increase over the past 45 years in the contribution to winter rainfall from heavy precipitation events; in summer all regions except NE England and N Scotland show decreases; and,
- Sea-surface temperatures around the UK coast have risen over the past three decades by about 0.7 ºC; sea-level around the UK rose by about 1mm yr^{-1} in the 20th century, corrected for land movement (the rate for the 1990s and 2000s has been higher than this).

The UKCP09 Briefing report (Jenkins et al., 2009) indicates that (under the medium emissions scenario), by the end of the century:

- All areas of the UK will become warmer; more so in summer than in winter. Changes in summer mean temperatures will be greatest in parts of southern England (up to 4.2ºC) and least on the Scottish islands (just over 2.5ºC);
- Estimates of annual precipitation amounts show very little change. The biggest differences would occur in winter precipitation, with increases up to +33%, along the western side of the UK, but with small decreases over parts of the Scottish Highlands. The biggest changes in precipitation in summer, down by about 40%, would occur in parts of the far south of England; changes close to zero would occur in parts of northern Scotland; and,
- It is very unlikely that an abrupt change to the Atlantic Ocean Circulation (Gulf Stream) will take place.

The report also contains UK sea-level data: including predictions of sea-level rise to the end of the 21st century (relative to 1980-1999) and excluding 'land ice melt' and 'UK absolute time mean sea-level change over the 21st century' (relative to 1980 to 1999, but including ice melt). The projections suggest:

- The range of absolute sea-level rise around the UK (before land movements are included) is projected to be between 12 and 76 cm for the period 1990 to 2095 (for the Medium emissions scenario); and,

- Taking vertical land movement into account gives slightly larger relative sea-level rise projections in southern Britain, where land is subsiding; somewhat lower increases in relative sea-level are indicated for the north. Isostatic movements are expected to typically be between –10 and +10 cm.

4.5.6 UK Ice Sheet Models

It is clear, both from the modelling of future short-term British climate (as undertaken by UKCIP), and of long-term climate (as undertaken by the BIOCLIM research), that the climate of the British Isles is becoming warmer and that the UK is unlikely to experience a new glacial episode until far into the future. Nevertheless, the BIOCLIM models do indicate that the ‘normal’ Milankovich cyclicity will return to dominate global climate, once the effects of anthropogenic warming have receded. The first major Northern Hemisphere glaciation is projected to occur between about 170 and 180 kyr AP (see Figure 26) and many future glacial episodes are projected to occur within the next 1 Ma. These episodes will have the largest impacts on the natural evolution of the geosphere and biosphere of Britain during the lifespan of a buried radioactive waste GDF. Rates and patterns of erosion and sediment transport, relative sea-levels, isostatic adjustments to the crust and the character of the ground water regime will all change dramatically. Consequently, because future patterns of ice accumulation are likely to mirror those of past glacial episodes, understanding the evolution of past British ice sheets will inform predictions of growth and decay of future ice sheets and their impact on the environment.

Parameterised models of the extent, thickness and evolution of ice cover of the British Isles are confined to the ice masses that formed during the last Devensian (OIS 2) glaciations, because much of the proxy data indicating the extent of former glacial episodes (such as the Wolstonian and Anglian) has been removed by subsequent erosion, particularly in the onshore areas. The models are of two main types:

- Models that reconstruct ranges of ice thicknesses and extent, variations in ice sheet morphology, and basal shear stresses of the BIIS during the Main Late Devensian glaciation, in which the ice sheet form is computed from a prescribed ice sheet margin (Figure 29).
- Dynamic models, which divide the British and Irish Ice Sheet into sectors and sub episodes (e.g. Younger Dryas/Loch Lomond Stadial) (Golledge et al., 2008; Hubbard, 1999; Hubbard et al., 2009). These 3D thermomechanical models use calculations of ice sheet rheology, climatic mean temperatures and precipitation patterns as baselines, and are driven by calibrated values of climatic variations from the North Greenland Ice Core Project (NGRIP) oxygen isotope curve (Rasmussen et al., 2006) and by the SPECMAP sea-level reconstruction (Ruddiman et al., 1989). Isostatic response to ice loading is computed using an elastic lithosphere/relaxed asthenosphere scheme.

The advantage of the ‘traditional’ ice thickness and extent models is that they provide contoured thicknesses for ice sheets, typically at their presumed maximum extent, which can provide baselines for GIA models that are used for reconstructions of past sea-levels and rates of crustal rebound that can be projected into the future (see below). Their difficulties are that most have been fitted to the presumed onshore extent of the BIIS at its maximum, and until recently most have taken relatively little account of the offshore evidence suggesting that the BIIS and the FIS were confluent and extended to the shelf edge well before the ‘traditional’ British LGM at c 22 ka. They have also failed to reproduce the full extent of ice within the

Irish Sea/Celtic Sea, which is now thought to have reached south to beyond the Scilly Isles (see Figure 22). This is largely due to the assumed basal shear stress values for the base of the ice. Figure 30a, for example was constructed assuming confluence between the BIIS and Scandinavian ice and used a basal shear stress of 100 kPa; 30c (based on limited ice extent in the North Sea) used a basal shear stress of 70 kPa on land and 30 kPa offshore. The widespread recognition of soft deformable beds, and fast-flowing ice streams, particularly in the offshore areas (see Figure 22) means that the traditional hemispherical 'domed' shape of dominantly 'land-based' ice sheets (such as the Laurentide ice sheet of north America) is not directly applicable to the form of the BIIS and that relatively thin ice lobes advanced and retreated rapidly when ice impinged beyond the present onshore areas (Boulton et al., 1977; Boulton et al., 1985; Boulton et al., 2002; Lambeck, 1991) (see Figure 30).

Most reconstructions of the form of the BIIS show that the ice blanketed even the most mountainous areas of the UK and, at its maximum extent, locally reached thicknesses >2 km. No comparable reconstructions are available for earlier glacial events but, the margin of the Anglian (OIS 12) ice extended further south than the Main Late Devensian glaciation limit in southern Britain, indicating that a greater thickness of ice was present over a wider area, and for a longer time, than it was during the Devensian.

The advantage of the dynamic models (Figure 31) is that sequential runs show the nucleation, growth and decay of the BIIS through time, replicating ice streaming, readvances and retreats of outlet glaciers and modelling of basal thermal conditions. They highlight the complexity of the system and bring into question the concept of an ice sheet maximum position, with concurrent advances and retreats of different portions of the ice front throughout much of the glacial episode. Many of the input parameters can be adjusted, but seldom is an 'optimal' maximum scenario presented. These models however explicitly consider glaciation from coalescing Scottish, Welsh and Irish accumulation centres alone (and in the case of the Loch Lomond Stadial confine its extent to the Scottish Highlands) without consideration of ice from elsewhere.

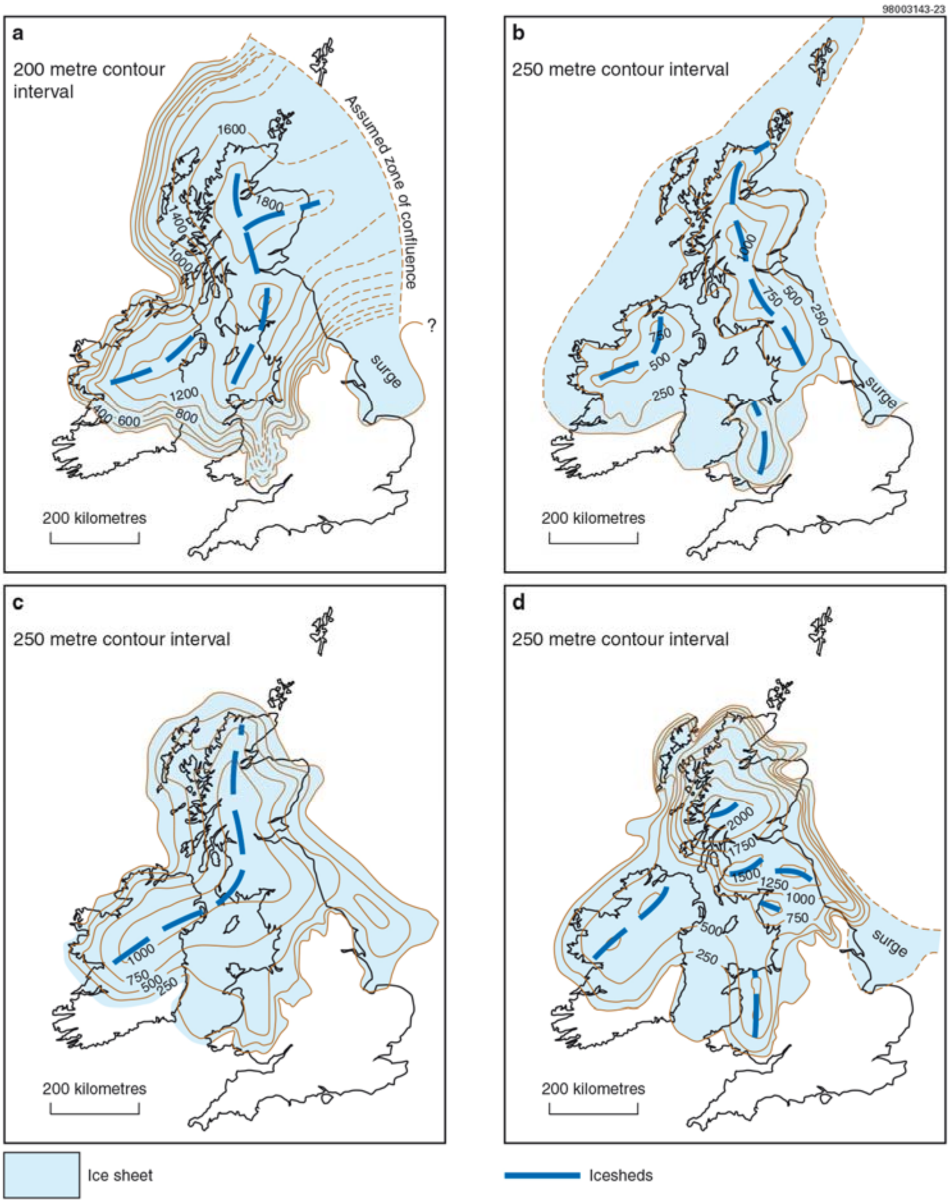


Figure 30. Models of the BIIS at its maximum extent (a) Boulton et al., 1977; (b) Boulton et al., 1985; (c) Boulton et al., 2002; (d) Lambeck, 1991. Modified after Merritt et al (2003).

The models of Boulton and Hagdorn (Boulton and Hagdorn, 2006) address some of these issues, including various scenarios for the confluence of the BIIS and FIS, identifying "slippery" patches, with low basal shear strengths, in offshore areas and giving contours on the ice sheet surface for some reconstructions. The authors are careful, however, to present the time intervals of the 'North sea confluence' models in terms of 'model time' rather than as specific time intervals.

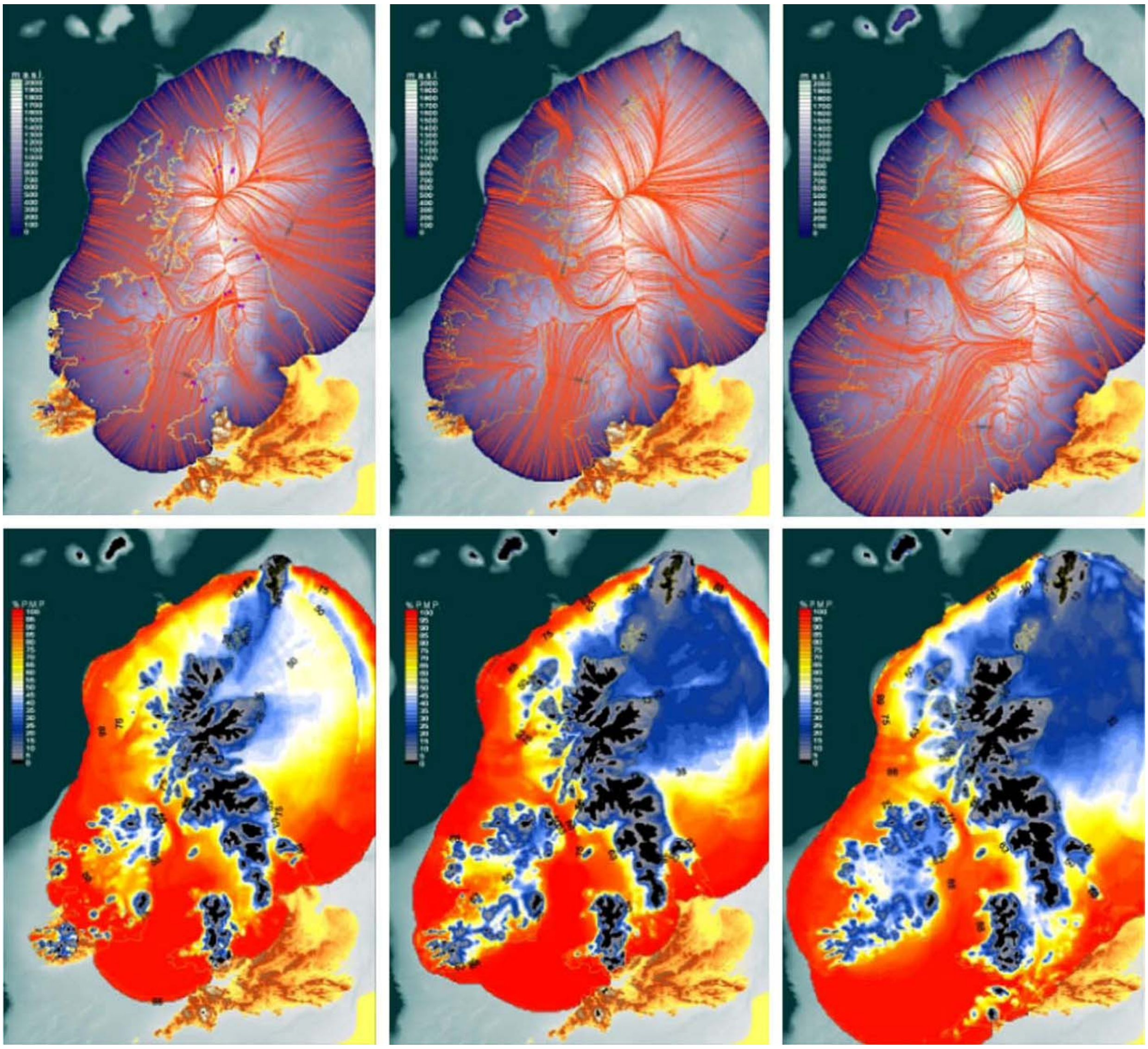


Figure 31. Examples of dynamic ice sheet models for 3 experimental conditions. Upper Row: Modelled ice sheet surface geometry and associated flowlines (red) for the time-slice corresponding to the maximum areal extent. Lower Row: Cumulative time that the ice sheet bed was at Pressure Melting Point (PMP) expressed as a percentage of the total simulation time. Persistent frozen basal conditions are given by black shading (%PMP < 2.5%) increasing to red shading (100%). Reprinted from Hubbard, A, Bradwell, T, Golledge, N, Hall, A, Patton, H, Sugden, D, Cooper, R and Stoker, M. 2009. Dynamic cycles, ice streams and their impact on the extent, chronology and deglaciation of the British–Irish ice sheet, Quaternary Science Reviews, 28, 758-776 with permission from Elsevier.

Models of the form and evolution of the last BIIS are important not only because they enable visualization of some of the most major environmental changes that affected Britain in the geologically recent past, but they are also the best indication of the pattern of any future glaciation that will occur. Ice-sheet models provide parameters for modelling rates and amounts of past and future uplift and depression of the land surface and accompanying changes in relative sea-level (see below).

4.5.7 Glacial Isostatic Adjustment models

These models link outputs from lithospheric and ice sheet modelling, calibrated by post-glacial sea-level data. As with ice sheet modelling, GIA models are confined to dealing with the effects of the last OIS 2 glaciation.

Most early estimates of past rates of vertical land movements are based on measurements of relative sea-level change during the Holocene (UKCIP02, 2002; Shennan, 1989; Shennan and Horton, 2002; Shennan et al., 2012). These indicate that rates have generally decreased with time, since deglaciation and throughout the Holocene, but that there was very considerable local variation in both the rate and amount of adjustment. In general, locations that were nearest to the centre of ice accumulation in the south-west Scottish Highlands were also closest to the centre of maximum isostatic depression. They have subsequently rebounded the most (Shennan et al; 2006a). Similarly, sites in southern Britain further from the centre of ice accumulation have rebounded least and been subjected to rising relative sea-levels throughout the Holocene. Figure 31 shows this trend continued during the last 1,000 years, with northern Britain rising 0.5-1.4 mm yr^{-1} and parts of southern Britain subsiding by 0.3 to 0.6 mm yr^{-1}.

In parts of Scotland, tilted raised shorelines terminating at ice margins indicate that during the early stages of deglaciation rates of isostatic rebound exceeded eustatic sea-level rise (Firth, 1989; Sissons et al., 1966). However, by the mid Holocene climatic optimum (c. 6 kyr BP), this situation had reversed, as indicated by the almost ubiquitous Holocene raised beaches around the present coastline.

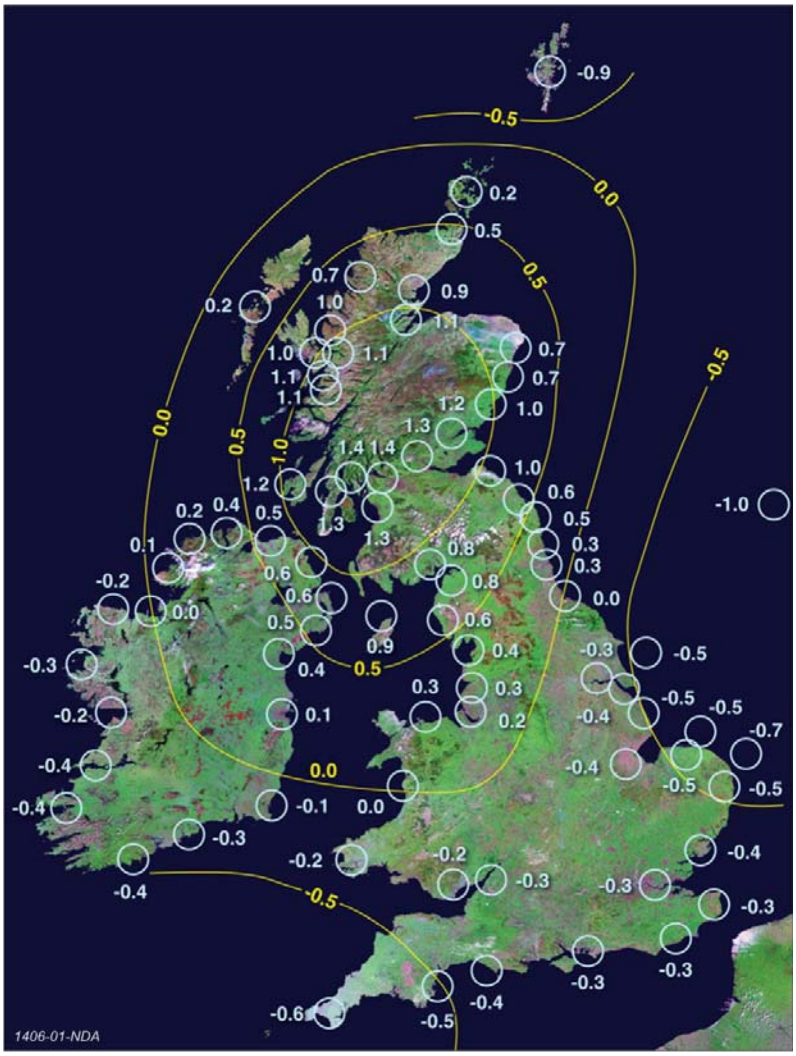


Figure 32. The mean rates of uplift and subsidence (mm yr^{-1}) during the last 1000 years i.e. the background geological trend, excluding any 20th century acceleration (NDA2010b).

Early attempts to directly link sea-level reconstructions to sequential models of BIIS evolution used minimalist 'traditional' ice thickness and extent models (Shennan et al., 2002), but better fits have now been attained, following research on Post-glacial relative sea-levels in Ireland (Brooks et al., 2008). These models (Figure 33) now take account of early confluence between the BIIS and FIS (Bradley et al., 2011), and their parting by 26 kyr BP, though the contours of ice thicknesses suggest that the ice sheet was thinner than indicated by models constrained by ice sheet rheology, flow rates and basal shear stress (Boulton and Hagdorn, 2006).

These ice sheet models have provided a good fit to the latest GIA models of recent isostatic adjustments (Figure 34), which include time-series GPS measurements from across the UK (Milne et al., 2006; Bradley et al., 2009). These show similar rates of isostatic adjustment to those provided by Late Holocene sea-level reconstructions (Figure 31).

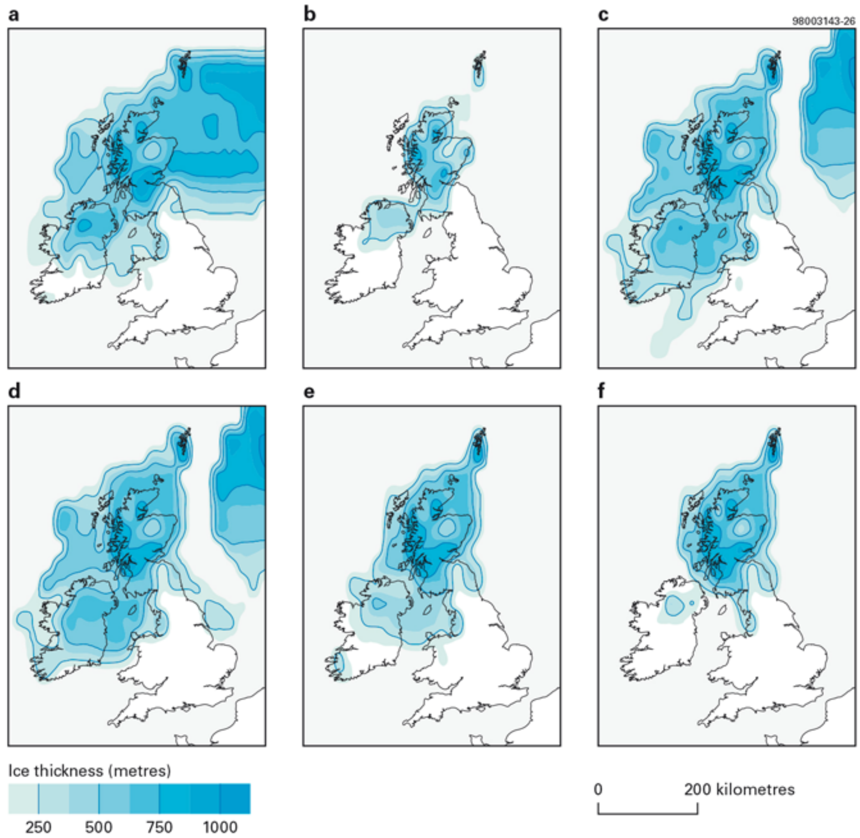


Figure 33. Examples of ice thickness models of the BIIS (a) at 32 kyr BP, (b) 26 kyr BP, (c) 24 kyr BP, (d) 21 kyr BP, (e) 20 kyr BP, 19 kyr BP, from (Bradley et al., 2011).

4.6 PAST AND FUTURE SEA-LEVELS

Sea-level changes have occurred throughout the Earth's history and their magnitude and timing are extremely variable. In the time frame relevant to the deep underground disposal of radioactive waste, eustatic changes are principally a consequence of thermal expansion and contraction of water in the oceans due to climate variations, and also to the extraction of water from the oceans during ice sheet growth and its return during ice sheet melting.

Evidence of high sea-levels in Britain during past interglacial periods is fragmentary. Probably the most complete record occurs in the coastal zone of Sussex and Hampshire where a succession of raised beaches, its chronology constrained by luminescence dating, is found. Raised beaches assigned to OIS 11, 9, 7 and 5e have all been recognised at elevations that range up to c. 50m above Ordnance Datum (AOD) (Bates et al., 2010). Elsewhere, OIS 7 and 5e raised beaches on the Gower in South Wales (Bowen, 1999c); an OIS 7 raised beach at Easington in County Durham (Davies et al., 2009) and an undated (presumed OIS 5e) raised beach, beneath till on Hoy, in Orkney (Sutherland, 1993) all indicate that relative sea-levels higher than at present, occurred in the coastal areas of the UK during the interglacial episodes that followed the Anglian (OIS 12) glaciation.

The most compelling evidence of low relative sea-levels during the onset of The Main Late Devensian glaciation is provided by imagery of the sea bed of the northern sector of the North Sea, where landforms such as moraines, glacial drainage and esker ridges, typical of terrestrial glacial environments, have been recognized extending to the edge of the continental shelf (Bradwell et al., 2008).

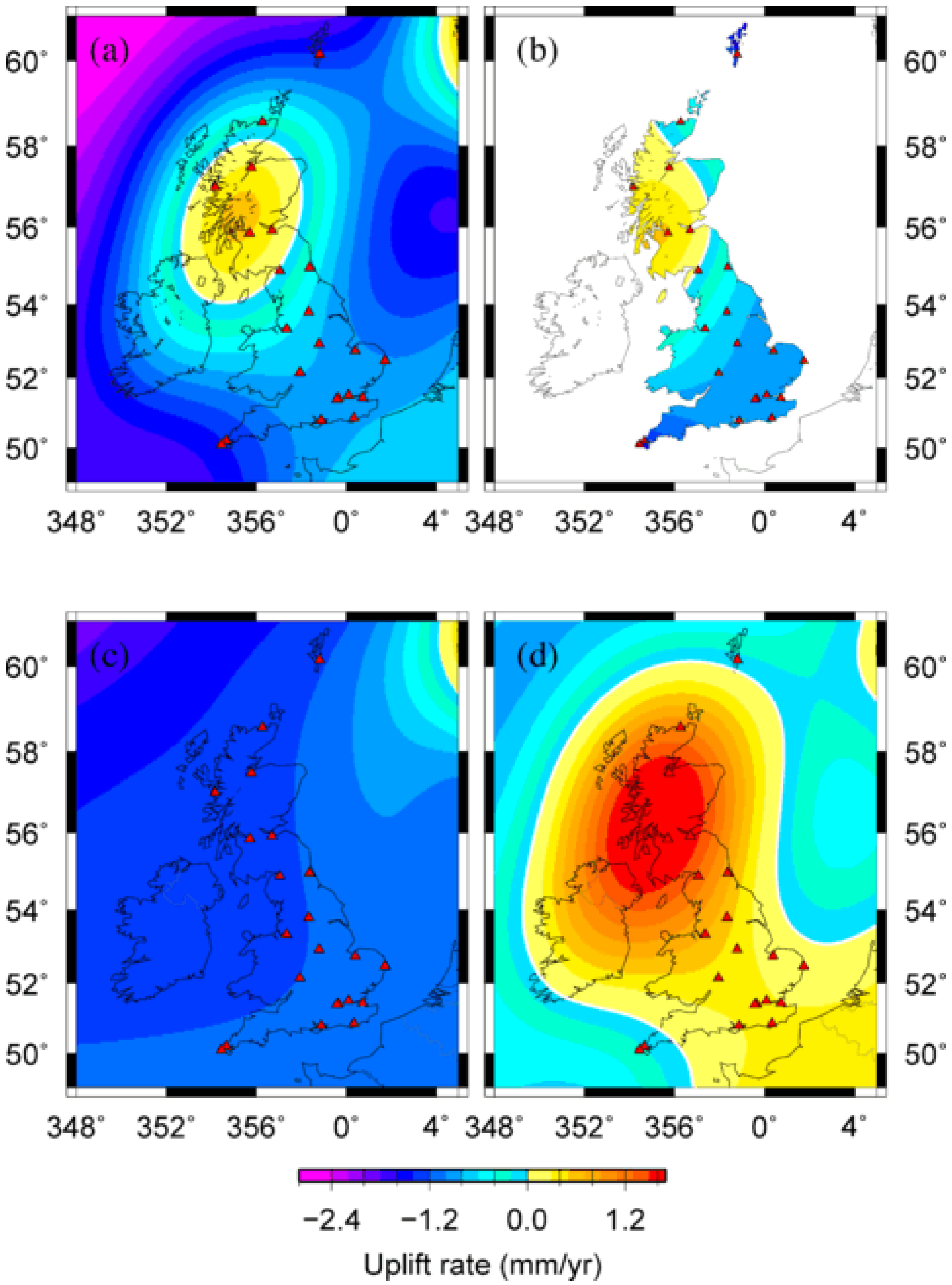


Figure 34 Predicted uplift rates for the UK from GIA modelling and GPS measurements (a) Predictions of uplift rate on a 5 km by 5 km grid for the reference ice model and an Earth viscosity model that, when combined with the reference ice model, provides a good fit to the regional sea-level database (Shennan et al., 2006a). This model adopts a 71 km lithosphere thickness, and a upper and lower mantle viscosity of 5 × 1020 Pa s and 1 × 1022 Pa s, respectively. The large positive uplift rates shown in the top right of the frame are associated with the deglaciation of the Fennoscandian component of the adopted ice model. (b) Same as (a) except that predicted velocity field sampled only at GPS site locations. (c) Component of total predicted signal (a) associated with non-local ice sheet loading. (d) Component of total predicted signal (a) associated with local (British–Irish) ice sheet and ocean loading. (This figure is a reprinted from Figure 2 in Bradley et al. (2009)).

The distribution of Late glacial raised shorelines in Scotland, mentioned above, indicate that relative sea-levels were locally more than 30 m above present levels. Their elevation decreases away from the centre of isostatic rebound in a similar manner to that of the more widespread and well documented Holocene shorelines which generally occur within or below about 10 m of present sea-level.

The modelling of sea-levels far into the future is a complex task that involves assessing past records of long-term global eustatic sea-level change and linking this to long term climate models. This has been done by Goodess *et al*., (2004), who undertook a study of potential changes in global sea-level over the next 150 ka. They used a method for estimating sea-level change from changes in Northern Hemisphere ice volume using a linear regression model. The model was developed using evidence of past sea-level as the predictand and the BIOCLIM LLN 2-D simulated ice volume as the predictor variable. The regression model was able to reproduce the observed sea-level changes over the last glacial-interglacial cycle, as indicated by calibration against proxy records of coral reef growth on the Huon Peninsula in Papua New Guinea and in Barbados (Figure 35). It did appear however, to underestimate sea-level fall, most notably during the LGM. The results from the model implied that the Antarctic ice sheets only made a small contribution to global sea-level change during the last 125 ka, although the result is not well constrained (see below). The regression model was applied to output from eight LLN 2-D simulations which incorporated natural and anthropogenic changes in atmospheric CO_2 concentrations, and one which only incorporated natural changes. This produced scenarios of eustatic global sea-level change, with 95% confidence limits, for the next 150 kyr years (Figure 36).

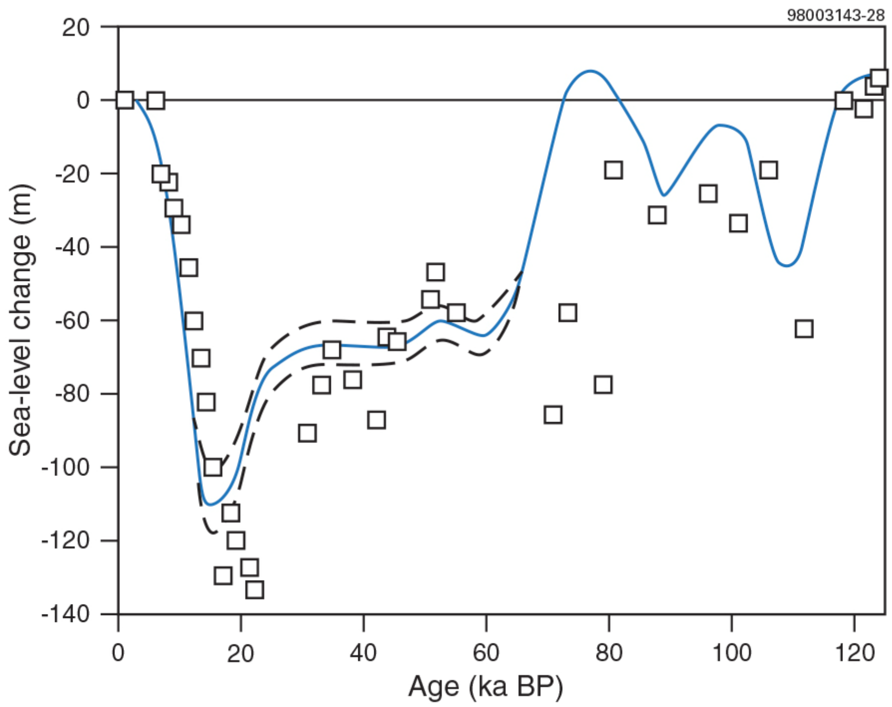


Figure 35 Predicted sea-level change (m) at 1 kyr intervals over the last 125 kyr (solid line) with 95% confidence limits (dashed lines), compared with observed (Huon Peninsula and Barbados) changes (squares) (Goodess et al., 2004).

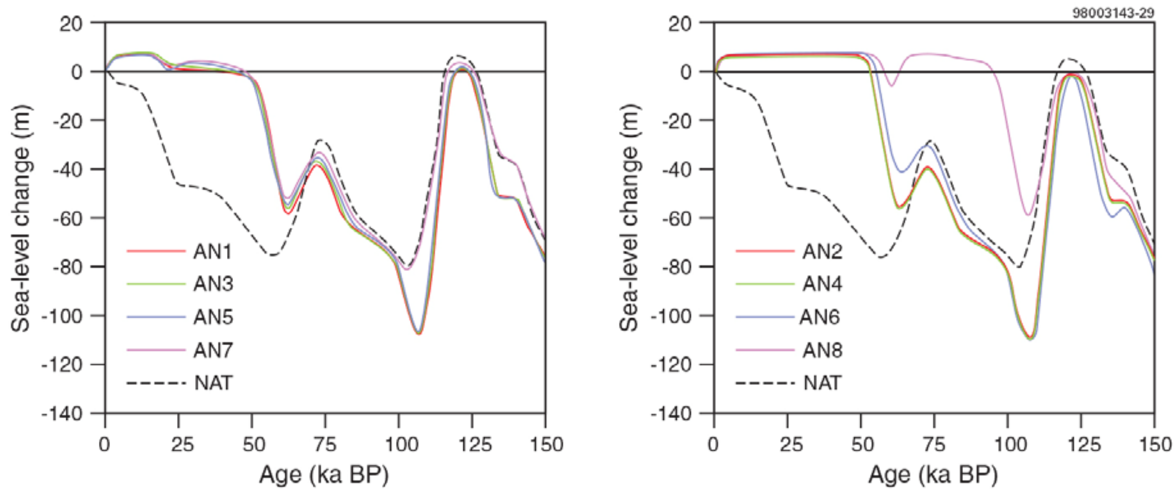


Figure 36 a): Future eustatic sea-level change (m) for low (AN1, 3, 5, 7) CO_2 forcing scenarios and the natural (NAT) simulation. b) high CO_2 forcing (AN 2, 4, 6, 8) (Goodess et al, 2004). "High" and "low" scenarios used were those of Sundquist (1990). The different scenarios assumed that the anthropogenic effects tailed off at 30 (AN1 & 2), 50 (AN3 & 4), 100 (AN5 & 6) and 150 ka AP(AN7 & 8). NAT assumes no anthropogenic effect.

For all of the CO_2 forcing scenarios sea-level rises (or is above present) for about the next 50 ka, after which it drops sharply coinciding with the beginning of modelled growth of Northern Hemisphere ice sheets. This culminates as falls of global sea-level of between c. 80 – c. 120 m at a glacial maximum that occurs between 110 and 125 kyr AP. This is followed by a rapid sea-level rise associated with widespread deglaciation. For the highest forcing scenario (AN8) the rapid fall in sea-level at c. 50 kyr AP is curtailed and only begins shortly before the glacial maximum. These models have subsequently been used to identify low-lying coastal areas around Britain (Figure 37) that would be susceptible to future sea-level change over the next 150 kyr (Hagdorn, 2003).

Figure 37 Coastal areas around Britain that would be susceptible to future sea-level change over the next 150 ka. Modified after Hagdorn (2003).

4.6.1 Global relative sea-level change in the 21st century

The IPCC 4 report (Solomon et al., 2007) (http://www.ipcc.ch/) projected century-end sea-levels using the Special Report on Emissions Scenarios, but did not include all possible sea-level rise drivers. The projections in the report of 0.18 - 0.59 m sea-level rise by the last decade of the century (compared with the average for 1980–99) are now considered to be conservative. Allison et al. (2009) concluded that sea-level rise by the end of the century is likely to be at least twice as large, with an upper limit of c.2.00 m; a view endorsed by the National Research Council of the National Academies (2010). There is the possibility that rapid change in glaciers and ice sheets, will greatly affect sea-level. Projecting partial deglaciation of the Greenland ice sheet, and Antarctic ice sheets, Rignot et al. (2011) propose a mid-Century sea-level rise of 0.32 m, whilst further melting of ice cover from these sources could contribute between 4.00 m to 6.00 m more to sea-levels over centuries to millennia. The most recent ensemble study of satellite imagery (Shepherd et al, 2012), published in November 2012 indicates that, since 1992, melting of the Greenland and Antarctic Ice Sheets has contributed, on average 0.59 ± 0.2 mm per year to the rise in global sea-level (equivalent to 11.2 ± 3.8 mm in 20 years. This accounts for some 20% of current sea-level rise (Kerr, 2012).

4.6.2 Predictions of relative sea-level change in Britain in the 21st century

The UKCP09 global mean sea-level projections were produced using results from the IPCC Fourth Assessment Report (Solomon et al., 2007). It gives an estimated range (5th to 95th percentile) for global sea-level increase of 18-59 cm between present day (assuming a1980-1999 baseline) and 2090-2099 (Millin, 2010). The report estimates that approximately 70% of the global sea-level rise over the 21st century will be due to thermal expansion, with the remainder due to the melting of glaciers and ice caps, with a combined contribution from the Greenland and Antarctic ice sheets.

The UKCP09 estimates for regional sea-level change around the UK use the estimates from the IPCC AR4 report, which used 16 atmosphere-ocean models (multi-model ensemble or MME) simulations, forced by the medium emissions scenario. These were then combined with estimates of the land ice melt component for the appropriate emissions scenario from the IPCC AR4. This gave a total (absolute) projected sea-level change for the UK for three scenarios over the 21st century, before consideration of land movement (Table 4).

Climate Scenarios	5th percentile	Central estimate	95th percentile
High emissions (A1F1)	15.4	45.6	75.8
Medium emissions (A1B)	13.1	36.9	60.7
Low emissions (B1)	11.6	29.8	48.0

Table 4. UK absolute time mean sea-level change (cm) over the 21st century, including ice melt, under three different scenarios, with 5th to 95th percentile confidence intervals. Descriptions of A1F1 and B1 can be found in table 2. Storyline A1B is as A1F1, except with a balance between fossil and non-fossil fuels sources). Data from Millin (2010).

Land movements, derived from the GIA and ice-loading models and GPS measurements (Shennan et al, 2006b; Milne et al., 2006; Bradley et al., 2009; Teferle et al., 2009) were then used to present estimates of future changes in relative sea-level under the three emissions scenarios across the UK. These were illustrated for four sites (London, Cardiff, Belfast and

Edinburgh) (Figure 38). All show similar rising trends for relative sea-level, but the rise is greatest for the southern sites (London and Cardiff) under all climate scenarios.

These trends suggest that much of the coastline of southern Britain will be at risk of shoreline erosion during the next one hundred years, with varying degrees of severity, dependent upon the future socio-economic scenarios used. Four examples, based on different 'Foresight Futures' published by The Office of Science and Technology (Evans et al., 2004) are illustrated below (Figure 39); all show severe risks of coastal erosion around The Wash, the East Anglian coast and the inner Severn Estuary.

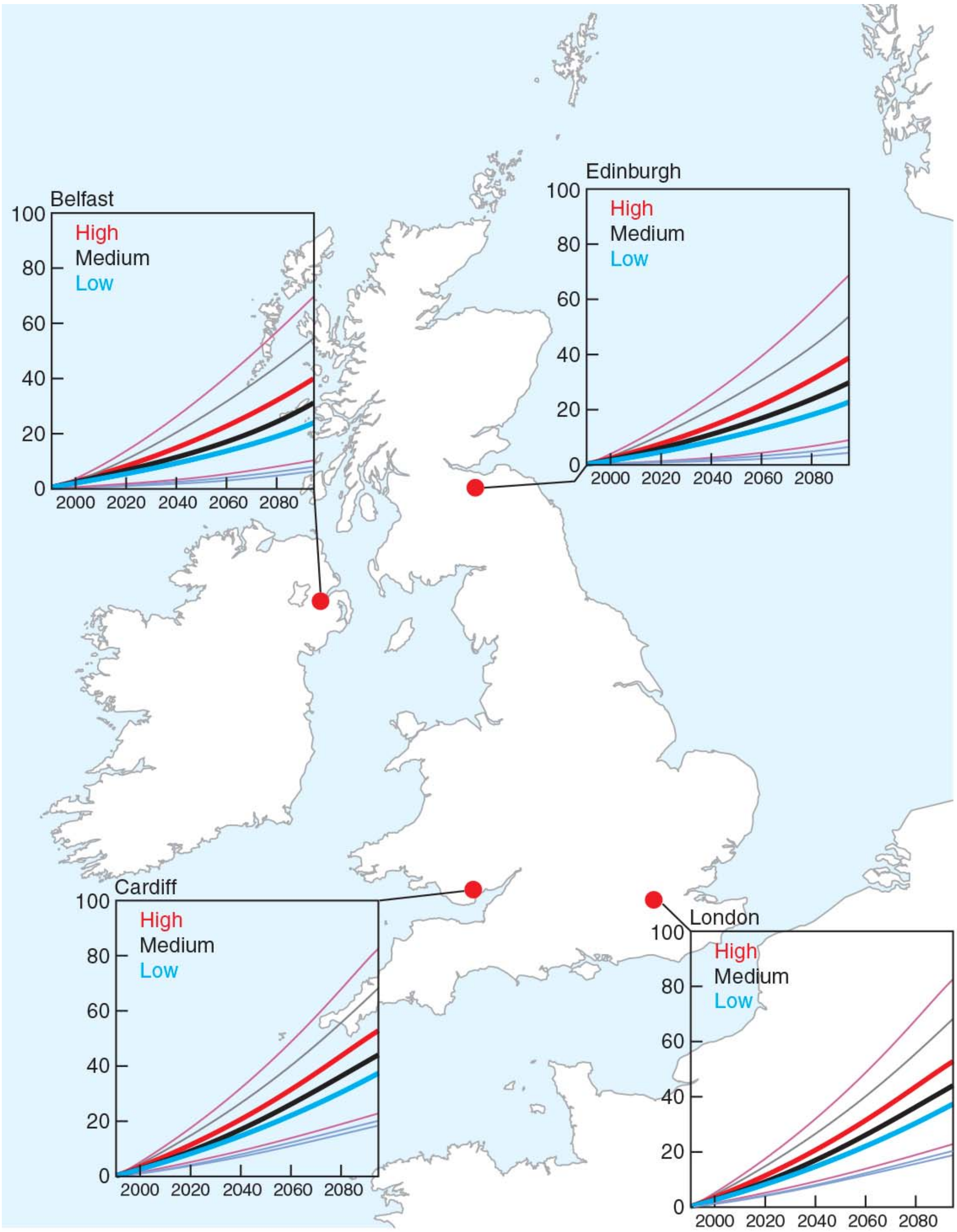


Figure 38 Central estimates for relative sea-level change across Britain under the High (thick red line), Medium (thick black line) and Low (thick blue line) scenarios of UKCIP09. Thin lines represent the 5th and 95th percentile limits of the range of uncertainty. Values are relative to 1990. Figure modified after Millin (2010).

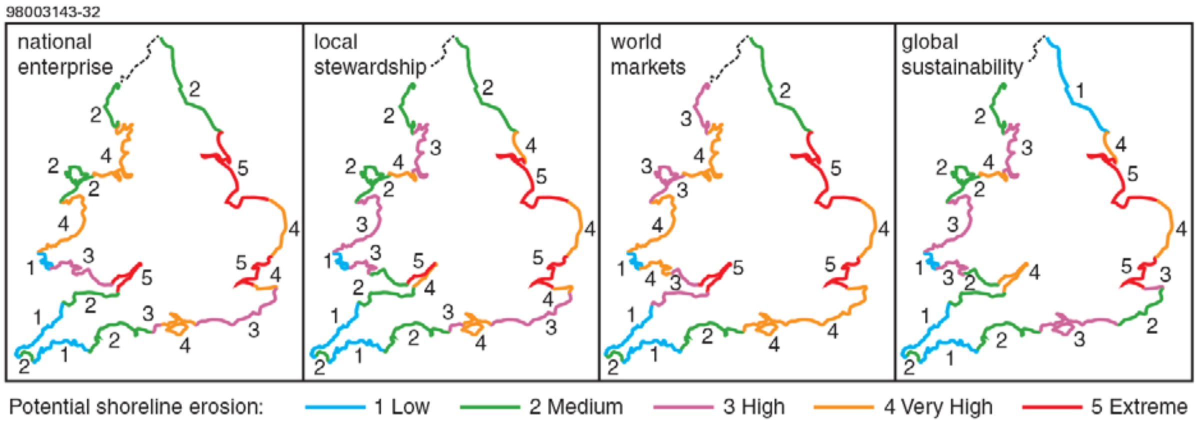


Figure 39 Regional differences in coastal erosion over the next 100 years under 4 socio-economic scenarios. Modified from Evans et al (2004); after Gibbard et al (2006).

4.7 PERMAFROST

As noted above it is likely that the climate in the UK will return to cold conditions at intervals over the next 1 M yr. During these cold climatic phases it is likely that many parts of the UK will experience permafrost conditions.

Quaternary geological studies indicate that during the last glacial cycle permafrost conditions may have prevailed over larger areas, over longer periods and at greater depths than previously believed. Since periods of permafrost can be expected in the UK, permafrost will be one of the few important climate scenarios for long-term repository safety and performance. In most of these areas permafrost will be followed eventually by periods characterised by a thick ice cover and the underlying permafrost layer will gradually diminish. However, for considerable periods of time, when global climate conditions are colder than the present day, areas of the UK will remain ice-free and therefore without the insulating influence of an ice cover permafrost will develop.

Available information from sites studied in NW Europe yields little quantitative evidence of past permafrost activity, whether geological, hydrogeological or hydrochemical. However, various completed or ongoing projects have, in a way or another, addressed permafrost and studies have also been conducted in sites presently experiencing permafrost conditions (i.e. sites in Canada, Greenland and Russia as natural analogues (e.g. Ruskeeniemi et al., 2002; Stotler et al. 2009a & b).

Significant amounts of modelling and experimental work have been conducted to establish freezing scenarios for crystalline bedrock, and lately, natural analogues have been exploited to verify if these scenarios are realized under natural conditions. Less is known about the impacts and features of permafrost in deep sedimentary environments. However, there is a vast amount of information on shallow systems, where the freezing is considered from geotechnical point of view. Relevant information is also generated during the exploration of methane ice resources from frozen sedimentary terrains. In addition, the concern related to the melting of permafrost in response to global warming has increased the interest in studying mountain and continental permafrost in various parts of the world.

Periglacial conditions with bedrock temperatures resulting in freezing may affect GDF performance through, for example, freezing of various parts of a GDF and surrounding rocks and through changes in groundwater flow and chemical composition. As a result, the safety functions of a GDF may be challenged by these processes. To date research has been addressed at site-specific studies of the amount of permafrost development under various climate scenarios (e.g. Hartikainen et al. 2010), and the possible consequences of freezing in the repository environments (e.g., Birgersson et al. 2010, SKB 2010, Vidstrand et al. 2010). Behaviour of buffer materials having a high porosity and water content is of special importance for the integrity of a GDF.

So far little attention has been directed to the effects of permafrost on sedimentary host rocks, including mudrocks. Johnson, (2002) gives a summary of the effect of permafrost, but only with respect to a severe reduction of dilution of released materials in the aquifer system in the overburden of the host rock. Wildenborg et al (2003) analysed the effects of glacial loading by a 1000 m thick ice sheet on the permeability characteristics, fluid flow rates and the associated migration of radionuclides both within and out of Tertiary clays in the Netherlands. It was found that glacial loading causes the expulsion of pore water from deeply buried clay deposits into adjoining aquifers. The rates and duration of the consolidation-driven outflow of water from the clay deposit were found to be very sensitive to the permeability of the clay and the dynamics of the advancing ice sheet.

For disposal in clay, there is a need to assess mechanisms that may impair the safety function of the clay layer. These include mechanisms that may increase diffusion (the dominant transport mechanism for radionuclides in clay) or decrease sorption to the clay. Effects of extreme hydraulic gradients related to an ice sheet margin, diluted glacial melt waters and increased salinity due to a propagating freezing front have not been analysed with respect to the clay host rock.

Current, temperate, conditions and their importance for subsurface conditions such as groundwater flow and groundwater composition are quite well known. The largest departure from the current situation (rock stress, pore pressures and groundwater flow) may be expected during periods of glaciation, and consequently investigations of the subsurface impact of climate change have to date focused on glacial conditions (ice-sheet influences). As a result of this history, the detailed effects of permafrost and related processes, which may have persisted for much greater periods of time, are less well known.

Based on the palaeogeographical extent of past Quaternary glacial and periglacial environments and associated permafrost depths (Delisle et al., 2003; Grassmann et al., 2010), future permafrost development may affect potential GDF locations in the UK. Direct impacts on a GDF, including possible damage to the EBS, may occur at locations where deep permafrost develops. However, even if repository depth is greater than the zone likely to be directly affected by permafrost development, impacts on the host rock and indirect effects such as brine formation and migration, intrusion of freshwater from melting permafrost or gas hydrate (formed beneath the permafrost layer (see below and Rochelle and Long, 2009), and cryogenic pore pressure changes associated with volume change during the water-ice phase transition may affect the integrity of the geological barrier.

Potential features and processes that may influence GDF behaviour, but for which there is currently little information, include:

- Changes in porewater geochemistry within a GDF at sub-zero temperatures during cycles of freezing (increased salinity) and thawing (freshwater penetration from melting permafrost ice or gas hydrates within the EBS or in the host rock;
- The effect of geochemical changes in porewater chemistry on EBS materials, which may alter and/or compromise barrier performance leading to an alteration of the hydraulic, swelling and sealing behaviour of bentonite and the long-term stability of cement and concrete materials;

- Deformation and changes to the fabric of the bentonite or cement containment materials by the formation of localized ice or gas hydrate wedges/lenses, and their subsequent melting during cyclic freezing and thawing, which may change the THM properties and create void spaces and local permeable flow paths; and,
- The production of carbon dioxide and methane as a result of the degradation of organic wastes within a cementitious low-level/intermediate-level (L/ILW) waste repository may ultimately lead to the formation of gas hydrates (methane hydrate, carbon dioxide hydrate) in the host rock or within the repository if deep permafrost developed. Subsequent warming and release of methane and carbon dioxide could have important impacts on the structure and fabric of EBS materials, resulting in changes in their thermo-hydro-mechanical-chemical (THMC) behaviour.

It is anticipated that permafrost conditions may affect safety-related components present in a GDF and safety functions assigned to the EBS and/or the host rock. These include:

- It is not well established if, and to what extent, bentonite seals and buffers, which are part of some repository designs, could be affected by permafrost-induced processes;
- The effects of permafrost conditions on radionuclide retention properties of clay host rock have not yet been investigated thoroughly;
- Permafrost may also affect the migration rates of released radionuclides through overlying stagnant aquifers. Freeze-thaw cycles could lead to alternating periods of accumulation and enhanced transport of radionuclides through the aquifers, which in turn could lead to periods of reduced and of increased exposures in comparison with the long term average; and,
- thermo-hydro-mechanical-chemical and biological (THMCB) properties of various repository and overburden lithologies are well understood for ambient and elevated temperatures but this is not necessarily the case for cold temperatures.

The presence of microbial communities in a range of extreme and subsurface environments is now well recognised (e.g. Lin et al., 2006; Roussel et al. 2008) and this includes groundwaters in and below the base of thick permafrost (e.g. Onstott et al, 2009). The impacts of microbial processes on the geological containment of radioactive waste is also long recognised (e.g. West and McKinley, 2002). As a result of these studies, it is concluded that microbial activity in any geological environment is only controlled by the availability of water, sufficient physical pore space and enough energy and nutrients to maintain metabolism over very long periods (thousands of years to million year timescales). Consequently, a permafrost environment cannot be assumed to be sterile and the presence of a repository will provide a potential source of nutrients and energy available for microbial use. This will be particularly the case for cementitious L/ILW repositories where a significant amount of organic waste will be present, and microbial activity may significantly influence the geochemical environment of the repository host rock and the EBS.

Current understanding of the hydromechanical, geochemical and multi-phase flow behaviour of compact bentonite, cement backfill materials and clay based host rocks is based predominantly on laboratory and field-scale observations performed at ambient or elevated temperatures (generally >20°C). There is a paucity of data for sub-zero environments associated with the development of permafrost conditions which limits model development and calibration (both conceptual and numerical) and represents a major uncertainty in the quantitative treatment of permafrost in repository performance assessment. Previous research has focussed on the evolution of individual parameters to freeze-thaw action such as swelling pressure (Birgersson et al, 2010, Birgersson et al., 2008, Schatz et al., 2010), hydraulic conductivity (Schatz et al., 2010) or the performance of non-buffer geomaterials (Thomas et al., 2009).

To date, research on the long-term stability of cementitious repository materials has focussed on the evolution of the cement systems as a result of the breakdown of wastes and interaction of the materials within contemporary groundwater systems. No attention has been given to the performance of cementitous materials when exposed to cyclic sub-zero conditions or to changing groundwater chemistry resulting from repeated freeze-thaw cycles. However, at low temperatures highly hydrated cement phases such as thaumasite may be stabilised. This has a high molar volume and may have significant impact on the physical properties of the cement.

4.7.1 Gas hydrates

A desk based scoping study was carried out in 2008 (Rochelle and Long; 2009) to establish if permafrost conditions possible in the UK under future cold climate conditions could lead to the formation of CO_2 and methane hydrates within the geosphere at depths relevant to a GDF. The study makes no assumptions about relative timing of GDF evolution and onset of cold climatic conditions. No assessment was made about whether or not CO_2 and/or methane would still be produced in a GDF and, if so, in what quantities, but it is assumed that both these gases would still be generated. Depending on degradation and corrosion rates and processes, such as biogenic production of methane from hydrogen these gases may still be present in and around a GDF when cold climatic conditions next occur in the UK.

Gas hydrates are a group of crystalline, ice-like solids that are generally stable under elevated pressures and lower temperatures. Many different gases form hydrates, methane hydrate is by far the most common hydrate on Earth, being widely distributed within polar regions and along deepwater continental margins where high pressure, low temperature environments coexist with a supply of gas. Hydrate distribution has changed over the recent geologic past, in particular as a result of waxing and waning glacial cycles. It is possible therefore, that future cold climate conditions within the UK geosphere might favour the formation of hydrate phases, and that they might form in the vicinity of a GDF.

There is a range of factors that affect hydrate stability, and not all of these were included within the scoping study. The variables that were considered were pressure, temperature and salinity, and it was assumed that sufficient amounts of gas were available in solution for hydrate to form and that pore size had no effect on hydrate stability. Thermodynamic equilibrium was assumed, and time-dependent factors such as the rate of hydrate formation, or rate of heat input were not taken into account.

The main results of the scoping study are:

- Neither CO_2 or methane hydrates are stable within the UK onshore geosphere under present day conditions (assumed average surface temperature of 10°C and a geothermal gradient of 25°C/km);
- For average surface temperatures below 0°C (permafrost conditions) CO_2 hydrate becomes stable in the presence of dilute groundwater at a depth of about 300 m (with an assumed geothermal gradient of 25°C/km and pressure controlled by hydrostatic head). Methane hydrate requires colder conditions, and an average

surface temperature of -5°C would be needed to first stabilise it at about 400 m depth under similar conditions;

- Under likely permafrost conditions (average surface temperature of -10°C), both CO_2 and methane hydrates could be stable at likely (400-800 m) GDF depths. This is particularly the case in the presence of dilute groundwaters, but they could also form in the presence of relatively saline groundwaters;
- A 1 km thick ice sheet having a basal temperature of 0°C would stabilise CO_2 hydrate beneath it in the presence of a range of fluid salinities. The CO_2 hydrate stability zone is limited however, and would not penetrate below a depth of about 400 m. The situation is similar for ice sheets of 500 to 1500 m thickness, however it changes if a basal ice sheet temperature of -5°C is assumed. In this case CO_2 hydrate could be stable at depths up to about 600 m; and,
- A 1 km thick ice sheet having a basal temperature of 0°C would also stabilise methane hydrate beneath it in the presence of a range of fluid salinities to about 600 m, and the more dilute of these would facilitate methane hydrate stability. Thicker or thinner ice sheets would impart slightly deeper or shallower hydrate stability zones respectively. However, there are larger changes if a basal ice sheet temperature of -5°C is assumed. In this case methane hydrate could be stable in the presence of relatively dilute groundwaters to depths approaching 1000 m.

Given that there appears to be potential for hydrate formation within and around a GDF, it may be appropriate to undertake a more quantitative assessment of the processes involved that includes timing and duration of future cold climatic phases and an assessment of whether any gases may be available for hydrate formation. If timing of future cold phases and related availability of gases suggest that hydrate formation may occur possible areas for study include:

- Establish a realistic thermal model for different areas of the UK covering several glacial cycles;
- Produce a realistic model for changing pressure gradients through the relevant parts of the geosphere caused by ice loading;
- Investigate likely gas compositions and production rates within a GDF;
- Ascertain whether hydrate formation and disassociation will affect the physical properties of a GDF; and,
- Assess the potential for, and impact of, episodic gas production as a result of hydrate breakdown.

Investigations such as those above should lead to an improved understanding of the potential for hydrate formation within a repository and the surrounding geosphere. Ultimately, this should allow an assessment of whether hydrate formation will be a problem, possible benefit, or of neutral consequence for the safety case of a repository.

4.8 OTHER IMPACTS OF GLACIATION ON THE ENVIRONMENT

There are a number of ways in which the geosphere would change as a result of a glacial-deglacial cycle:

- Glaciation is accompanied by erosion, sediment transport and dispersal, which would remove some near-surface material and redistribute much of the remainder;
- Loading and subsequent unloading of the Earth's crust as an ice sheet advances and then retreats causes subsidence of the Earth's crust followed by post-glacial uplift. This isostatic adjustment is a long-lived process. Stress changes caused by ice loading and unloading can induce fracturing and faulting in rocks, which may change some rock properties and cause displacements in the rock at depth;
- Glaciation may cause groundwater composition and groundwater flow patterns to change.

Deposits laid down by glaciers and by their meltwaters, form the principal substrates upon which the present landscape and soils of the UK have developed. Deposition by glacial action however, is of less direct potential impact on a buried GDF than the possibility of disturbance by deep glacial erosion. The extent and depth of glacial erosion is dependent on the speed of flow and thickness of the ice, as well as the nature of its bed. Less erosion takes places when a glacier is frozen to its bed, which generally occurs when the ice is thin, or at ice-sheds. Conversely, enhanced rates of erosion occur within areas of fast flowing ice, such as ice streams, which 'drain' the more stable areas of the ice sheet. The basal thermal regime of glaciers also affects what happens at their base and beneath their beds. Wet-based glaciers slide and erode their bed. They have a water layer at their base that may be forced into the rocks beneath, recharging the groundwater with fresh, oxygenated water, perhaps at elevated hydrostatic heads. On the other hand, cold-based glaciers are frozen to their beds and little glacial erosion occurs.

These zones of 'cold-based' slow moving ice and 'warm based' fast flowing ice will migrate as the ice sheet evolves (see Figure 30), but reconstructions of the OIS 2 BIIS show that ice-streaming affected many coastal zones sites. For example, an initial study of potential future glacial erosion at the Hunterston site on the coast of Ayrshire indicated a high likelihood of topographic ice sheet-flow focusing ('ice streaming') and that future glacial erosion rates are very likely to be some of the highest experienced in Scotland with an erosion depth range of 2.65 mm to 5.30 mm per year postulated. Locally, glacial erosion rates can vary considerably, possibly by one or two orders of magnitude (Hallet et al., 1996), depending on the nature of the substrate, the size of the source area for the glacier, its distance from its source, the lithology and grain size of material in the source area, ice flow velocity and duration of the ice flow event. For example, plucking may preferentially remove pre-fractured rock quickly (>1000 mm yr-1) in areas where subglacial stress fields fluctuate; whereas abrasion acts typically much more slowly (<1 mm yr^{-1}) even where basal pressures (and debris content within the ice) are high and concentrated. Glacial meltwater erosion can be highly variable in time and space and less predictable than plucking and abrasion – focusing mainly, although not exclusively, in topographic lows and along pre-existing channels. Rates of meltwater erosion can range from 1 to 1000 mm yr^{-1}, depending on the hardness and cohesive strength of the substrate, but may greatly exceed these values over short periods in exceptional circumstances (>10,000 mm yr^{-1}).

The advance and retreat of an ice sheet will affect the crustal stress field and may cause rocks to deform. The evolution of these stresses in the geosphere can be examined through modelling. For example, the impact of future glaciations on the crustal stress field has been modelled in Swedish studies in order to understand the natural evolution of the geosphere at the potential GDF sites. Peak impact on geosphere stress and rock deformation is predicted to occur as the margins of the ice sheet pass over the site. This generates a 'forebulge', or area of uplift, immediately in front of the advancing or retreating ice sheet. The properties that control the deformation and stress change are related to areal extent and thickness of the ice sheet and to the slope of the ice sheet margin. This change in confining stress, coupled with changing anisotropy of the stress field, has the potential to induce some opening or closing of pre-existing mechanical discontinuities in the rocks. During a period of glaciation when a regional ice sheet covers the land surface, it is expected that faults and fractures in the rocks would be stabilised. However, the changing stress field at a site as an ice sheet advances or retreats may reactivate pre-existing faults and cause seismic activity (See Chapter 3).

As the hydrogeological properties of higher strength rocks are controlled by the properties of these fractures, in particular their apertures, this has the potential to affect the permeability of the rock and hence groundwater flow. A future glacial episode would also have effects upon the pattern of groundwater flow, the amount of groundwater recharge and groundwater chemistry. During warm (interglacial) periods, the amount and distribution of recharge to the groundwater table would be controlled by factors such as the amount of precipitation and the topography of the land surface. During a future glacial or periglacial climate state, the exchange of water between the biosphere and the geosphere would be largely controlled by the extent and nature of the ice and permafrost cover (see Chapter 4). During an ice sheet glaciation which blankets the underlying landscape, it is the form and elevation of the ice sheet surface that provides the dominant topographic control over amounts, types and distribution of precipitation, but dynamic ice sheet models indicate the possibilities of rapid transitions, both in time and space, between glacial and periglacial environments

Under glacial conditions, the extent of groundwater recharge would be largely controlled by the hydraulic conditions at the base of the ice sheet, where it comes into contact with its bed. In some areas, the ice sheet would be expected to be frozen to its base; little groundwater recharge can occur in such areas and very little erosion occurs. Elsewhere water from 'basal melting' of the ice sheet would enter the water table where it would be joined by surface runoff percolating to the base of the ice (NDA, 2010a).

Dynamic ice sheet models of previous ice sheets can show where, when and for how long cold-based ice would be expected to occur (see Figure 30). It is notable that these conditions typify ice sheds or ice divides, principally located in the uplands on the northern and western sides of Britain (see Figure 29). Since the last major ice sheet glaciations of the British Isles, whose extents have been widely agreed (OIS 2 and OIS 12) had relatively similar extents, it is probable that future ice sheet glaciations may produce similar distributions of warm and cold-based ice and similar patterns of groundwater recharge.

4.9 CLIMATE CHANGE AND GLACIATIONS: POTENTIAL IMPACTS ON A GDF OVER THE NEXT ONE MILLION YEARS

The heterogeneous pattern of glacial erosion/mass transport is a problem that means values of 'averaged or mean erosion rates for a terrain over a glacial-interglacial cycle' are relatively meaningless. The areas where deep erosion occurs will be localised and will be dependent on where active ice streams, major glacial meltwater drainage routes and major fluvo-glacial outflow incisions occur. In, and adjacent to, upland areas these are likely to be mainly controlled by the location of existing valleys which will control where future glacial and fluvial erosion is likely to occur. The eventual depth that the valleys and channels attain will depend on factors such as ice thickness and sea level, but all depth increases are probably in the 200 m rather than 1 km depth range. Within most of the UK landmass few incisions exceed 200 m in depth after multiple past glaciations; none are likely to reach 1 km depth below the present ground surface of any exiting valley unless many future glacial episodes occur.

In lowland ice sheet areas the depth of glacial erosion is likely to be of the order of a few 10's of metres. However, this lowering of the land surface will be less focussed and more widespread than in upland areas. There will also be areas of till and glaciofluvial outwash sand and gravel deposition over new or existing land surfaces. There is also the possibility of excavation of 'glacial rafts' of bedrock 100's of metres long and perhaps 50 to 100 m wide (for example the chalk rafts of the Norfolk coast). Large slabs of largely intact bedrock of this type are known to be moved by ice sheets, especially in areas close to their margins. In lowland areas, erosion by sub-glacial meltwater streams will not necessarily be governed by ground surface topography and the development of erosion channels by these streams will be controlled by the location of meltwater outlets and also by the elevation of driving heads. The latter can be raised to the level of the ice surface by the drainage of supraglacial streams to the glacier bed through moulins. Such channels may be eroded into the ground surface to a depth of a few 10's of metres and are frequently filled with glaciofluvial sediments and/or tills on retreat of the ice.

A GDF at a depth of 200 to 300 m sited in an area where valley glaciers or sub- and pro-glacial streams may be focused by current topography may be exposed or nearly exhumed by these processes over the 1 million year time frame.

As ice sheets grow and decay and climatic conditions, particularly temperature and precipitation vary, the thermal regime of the ice can change from cold to warm based and vice versa. This means that erosion rates can vary, both in extent and intensity thorough time, at any given location. The situation is slightly more complicated however, because the erosive power of glacier ice is dependent not only on its thickness, but also on the gradient of the ground over which it flows and the hydraulic conditions at its bed. Thus for a given substrate, glacial erosion will be at its most effective when the 'Goldilocks' conditions of gradient, ice thickness and water at the glacier bed are sufficient to lubricate the bed, but not so little as to inhibit flow or (in the case of hydraulic pressure beneath the ice) so great that the ice mass becomes largely detached from its bed.

The combination of topography (precipitation shadow) or elevation (mountain tops are covered by less ice in an ice sheet glaciation than pre-existing valleys) will also influence ice thickness and glacier type. In upland glacial systems frost shattering of areas not covered in ice may be as effective at lowering the mountain tops as the glaciers are in deepening their beds. It also provides debris that can be entrained in the ice to give glaciers their 'teeth'. Permafrost may extend beneath cold based glaciers, as well as beyond the margins of both warm and cold based ice masses, depressing groundwater recharge. If deep permafrost forms it may affect the engineering properties of 'soft' rocks and could lead to the development of new fracture pathways, perhaps to the surface, in more brittle formations. It also affects groundwater recharge and discharge amounts and positions. The formation of ice changes the chemistry of the remaining liquid phase, concentrating dissolved salts. Permafrost could also affect the engineered elements of a GDF in similar ways, in particular the properties of clay and cement based backfill/buffer materials may be permanently or temporarily changed by permafrost conditions. Because halite is a 'dry' rock it will not be affected by permafrost.

Depending on a number of factors, including the timing of post closure and the depth of ground freezing relative to a GDF, it is theoretically possible that permafrost conditions could lead to the formation of methane hydrates and carbon dioxide ices in the vicinity of a GDF which could be released quickly on warming. Evidence is that ice ages end quickly (perhaps only over tens of years) but often diachronously however warming would in geological terms be ‘quick’.

There are two main areas within which permafrost is potentially important for a GDF and for which research to develop understanding may be appropriate:

- The evaluation of the potential significance and impact of permafrost and related processes within the host rock at GDF depth and within surrounding geology (noting that the safety case for a repository relies on engineered and natural barriers working in complement in the post-closure period to deliver the necessary level of isolation and containment). In particular, a number of factors that may influence the groundwater chemistry and/or the migration of radionuclides are:
    - Cyclical freezing and thawing;
    - Increased groundwater salinity at freezing fronts;
    - Intrusion of freshwater during permafrost melting;
    - Formation and destabilization of gas hydrates; and,
    - Geomicrobiological influences under permafrost conditions.
- The geochemical and geophysical changes induced by permafrost in the host-rock environment need to be quantified to be able to assess their impact on the long term safety.

If relevant, the consequences of permafrost and related processes on the Excavation Damage Zone (EDZ) and the performance of the EBS, in particular with regard to their behaviour during transient periods when there may be high hydraulic, thermal or chemical gradients which could influence the long-term development and performance of repository components, need to be scoped. Most of these transient conditions in the near-field are driven by THMCB processes. These conditions can be weakly or strongly coupled to GDF safety and their impact for the different disposal concepts and scenarios needs to be assessed. Examples of processes that may affect the EBS include the swelling of bentonite, its interaction with steel and cement, gas generation and self-sealing of bentonite and clay host rocks.

Crustal stress increases caused by isostatic depression and unloading during repeated glacial-deglacial cycles may result in rock slope failures. This is especially possible in upland resistant mountainous areas when removal of ice support in deeply glacially, repeatedly occupied, incised valleys can lead to slope failure. While these rock slope failures are valley side phenomena that will have no effect at depths vertically below they can occur for perhaps a kilometre laterally from the valley sides in, say, the Lake District or the Highlands. They are often not recognised or are mapped as traditional landslips.

Changes of relative sea-level during glacial and interglacial stages could mean that a GDF site is further or nearer to the coast or even beneath the sea bed. While this may lead to coastal erosion or deposition around the surface part of a GDF only affecting a few 10’s of metres depth it will change the groundwater flow paths by changing the base level of discharges.

If global warming induced sea-level rises occur in the near future, that are as significant as some predictions suggest, then there is a risk of flooding during the active phases of GDF construction, waste emplacement and closure over a c. 150 year lifetime if the surface installations of a GDF are sited in a susceptible, low-lying coastal location. Post closure flooding of a GDF site will reduce groundwater driving heads and therefore probably reduce groundwater flows in the vicinity of a GDF.

Groundwater flows are likely to be significantly affected by glaciations and permafrost. Permafrost will freeze groundwater in-situ creating a rock mass that is effectively impermeable. This will lead to the diversion of any groundwater flow present around the frozen ground, perhaps changing groundwater flow paths and discharge locations, if groundwater flow is still active under these conditions. As noted above, the freezing process will concentrate any solutes present into a residual fluid phase which may be expelled as the ground freezes. These fluids will be saline and could affect a GDF using bentonite backfill. On thawing, the formerly frozen ground may have enhanced permeability because fractures may have been widened and pores enlarged during freezing that may not have returned to their original state when thawing occurred. This may lead to changes in groundwater flow paths and rates which, depending on the depth of a facility and of the frozen ground, may affect a GDF. Glacial processes will influence groundwater flows beneath and adjacent to ice sheets. The thickness of ice has potential to increase groundwater heads significantly, which in turn may force groundwater flows through the rock mass. This may in turn lead to changes in the groundwater flow pattern including higher heads, higher flow rates, longer and deeper flow paths. These in turn may lead to changes in the rock properties and groundwater chemistry that include dissolution and/or deposition of minerals in fractures and pore spaces, changes in solute concentrations and changes to the redox state. Ice cover or permafrost at discharge locations will divert groundwater flows and lead to the development of new discharge points, again potentially affecting groundwater behaviour. Such changes to groundwater flow are likely to be long term. Enhanced recharge by glacial melt water may lead to changes in redox conditions because of the introduction of oxygen-rich water. This is likely to be a near surface, perhaps 100 to 200 m, effect because of the buffering properties of the rock mass as a whole.

5 Weathering, Erosion and Palaeohydrogeology

5.1 INTRODUCTION

This chapter will expand upon the weathering and erosion aspects discussed briefly in the NDA Geosphere and Biosphere status reports (2010a and 2010b). A closed GDF, its host rocks and surrounding geology have the potential to be affected by surface erosion and weathering processes. There are many erosion and weathering processes identified in the geomorphological literature, but not all of these will have any effect on an appropriately located GDF. Table 4 outlines these geomorphological processes and their relevance for a closed GDF. Only those factors relevant for a GDF in the UK will be examined in greater detail in this chapter. Whereas a GDF may be located at depths up to a 1000 m the access infrastructure which will be backfilled will extend to the surface and this also needs to be considered. The major areas of weathering and erosion that will be considered within this chapter are:

- Glacial and fluvial weathering and erosional processes acting at the earth's surface;
- Subsurface weathering processes; and,
- Associated processes which can generate new weathering and erosive pathways leading to changes in ground or surface water movement (e.g. hydrofracturing, permafrost, river capture).

This chapter also covers diagenesis and related processes that alter the properties of rocks over time (diagenesis, in a broad sense in that the term is normally applied to sedimentary rocks but similar alteration processes occur in fractured metamorphic and igneous rocks) and palaeohydrogeology and related studies aimed at understanding past groundwater flow and chemistry as a means of predicting future behaviour.

5.2 CONVERSION OF WEATHERING RATES

Literature quoted values of denudation, weathering and erosion rates are presented with a variety of units, often according to scientific discipline. For example, denudation rates calculated using ^{10}Be or ^{26}Al are presented in metres per million years (m Myr^{-1}) whilst catchment studies measuring solute or suspended sediment discharges present estimates of erosion as tonnes per square kilometre per year (t km^{-2} yr^{-1}). For comparative purposes a very rough conversion of these rates can be made by assuming a bulk density of 2 t m^{-3} which will be within the accuracy of the determination of the discharges. Thus an erosion rate of 100 t km^{-2} yr^{-1} is equivalent to a denudation rate of 0.05mm yr^{-1} and a denudation rate of 0.01 mm yr^{-1} is equivalent to 20 t km^{-2} yr^{-1}.

5.3 CLIMATE SCENARIOS

Climate scenarios being considered are given in Chapter 4 and range from polar glacial to cool warm temperate conditions. By collecting data from present day weathering and erosion analogues of these climates, assessments can be made of the rate of likely weathering and erosion processes. However, the extent to which these processes may affect a GDF, particularly in relation to glacial processes will result from the temporal and spatial dynamics of the glaciation.

5.4 THE ROLE OF ISOSTATIC, EUSTATIC AND UPLIFT PROCESSES IN LANDSCAPE MODIFICATION

Erosion and weathering rates are heavily influenced by changes in altitude caused by uplift. Uplift is normally an interaction between tectonics and isostatic rebound and (i) creates new fresh material for weathering and erosion; (ii) causes river incision and (iii) modifies slope. Two major processes produce landscape uplift; these being (i) tectonic uplift which is connected to the production of mountain ranges (orogeny) and (ii) isostatic rebound after deglaciation.

Isostatic rebound causes uplift of the landscape through the unloading of ice during and after deglaciation. There is a general consensus that river incision approximates to isostatic uplift but lags slightly behind. Landforms associated with isostatic change include raised beaches and terraces. The current state of isostatic re-adjustment in the UK has been reported by Shennan et al. (2009) and shows that Scotland and northern England are currently rising by up to 1.4 mm yr^{-1} in central Scotland with a postulated ice sheet height maximum of 1500 m at 22000 y BP (Lambeck, 1995). However three foci of subsidence are located in southwest England, the southern North Sea and the Shetland Isles. These areas of subsidence demonstrate the effects of the ocean load on the Atlantic Basin and on the continental shelf and the glacial isostatic signal from far-field ice sheets including Fennoscandia. Rates of subsidence in southern UK can be up to -0.7 mm y^{-1} (Shennan et al, 2009).

Current rates of uplift in the UK are relatively low (Figure 32) and this reflects the thickness of ice during recent glaciations and its mid-plate geological setting. Greater uplift rates have been reported for Scandinavia, where uplift of 90 m has been estimated in central Sweden with a further estimated uplift of 90 m to be expected since the Fennoscandian ice sheet decomposed (Ekman and Mäkinen, 1996). This demonstrates the extent that glaciations will exert major controls on the amount of isostatic rebound experienced.

Landscape System	Geomorphological process	Low relevance	High relevance	Notes
Glacial	Moraine/ Drumlin / Esker deposition	x		Surface deposition of sediments may cover infrastructure
	Glacier over deepening		x	Potentially deep gouging of Earth's surface
	Buried valleys		x[1]	Glacial sediment filled valleys often formed by sub-glacial streams
	Flooded valleys		x[1]	Potentially deep gouging of Earth's surface
	Glacial-tectonic rafting		x[1]	Removal, transport and deposition of large pieces of bedrock
	Ice sheet movement		x	Erosion of land surface
	Hydrofracturing		x	Fracturing of rock due to sub-glacial/ice sheet water pressure
Periglacial	Patterned ground	x		Movement of surface due to freezing
	Active layer development		x	Can extend down to several metres and could affect top of infrastructure by freeze-thaw action
	Solifluction/gelifluction	x		Creation of head deposits on slopes
	Tallus / scree slopes	x		Creation of rock deposits on surface
	Permafrost		x	Can extend several hundred metres causing fracturing of rock and re-routing of ground water
Fluvial	River terrace formation	x		Formed as a result of isostatic rebound
	River incision		x[1]	Formed as a response to isotactic or tectonic processes
	Floodplain deposition	x		Sediment build up as a result of flooding
	River capture		x[1]	Caused by tectonics, ice damming
Terrestrial	Soil creep/erosion	x		Redistribution of Earth's surface through wind and water action.
	Land sliding	x		As a result of tectonic activity and increasing pore water pressure. GDF unlikely to be built in susceptible areas
	Debris flow	x		As a result of tectonic activity and increasing pore water pressure. GDF unlikely to be built in susceptible areas
	Karstification		x	GDF unlikely to be sited in a karst area but could form part of surrounding geology
	Denudation		x[1]	Continuous stripping of the Earth's surface through weathering, solution and erosion

Table 5: List of erosion/weathering processes and resulting geomorphological features and their relevance to GDF infrastructure.

x[1] - These processes may affect a GDF sited at shallower depths around 200 m below current ground levels in some parts of the UK.

5.5 TERRESTRIAL DENUDATION UNDER DIFFERENT CLIMATES

Climate has a fundamental role in denudation (erosion and weathering) because the majority of processes require, or are enhanced, by water. Tectonic activity in the form of mountain building is also important as the presence of new material generally increases weathering and erosion rates. Most surface denudation processes are unlikely to affect a GDF buried at 200 to 1000 m within the first one million years, but they have the potential to impact the backfilled structures and other infrastructure associated with a GDF. This section reviews the rates of denudation in different climates and by different methods.

5.5.1 Global estimates of bedrock denudation in different climates

Over the first one million years following closure of a GDF, a series of climates may be experienced in the UK with overall climate alternating between glacial and interglacial periods. For parts of the UK, the glacial periods will include glaciations (active ice cover) while unglaciated areas will experience permafrost conditions during these phases. During the intervening interglacial periods, the UK will experience cool to warm temperate climate. Relevant global denudation rates are presented in Table 6 where climate has been divided up into 4 categories according to previous definitions used by the NDA. These future climate states include Mediterranean type climate (warm temperate), temperate (current UK type climate), sub-arctic (boreal, periglacial, forest tundra) and polar (periglacial, permafrost and glacial) (NDA, 2010b).

Most of the rates of denudation erosion in Table 6 have been estimated using ^{10}Be or ^{26}Al, which are cosmogenic radionuclides produced within quartz grains via a process known as spallation. Their use has become widespread in dating and erosion studies. Other methods used include (i) Mass Balance calculations, (ii) Palaeosurface reconstruction and (iii) Apatite fission dating (AFT). Results show that the greatest denudation rates occur in orogenic environments such as the European Alps (Vernon et al. 2008) or the San Bernadino Mountains (Binnie et al. 2008) where bedrock lowering rates of 200 to 700 m Myr^{-1} and 70 to 1200 m Myr^{-1} have been estimated respectively. Typically, bedrock lowering rates in non-orogenic settings, such as the UK, are much less than 50m Myr^{-1}. With respect to different rock types, there does not appear to be strong relationships because other factors such as climate (particularly precipitation) would appear to be major determinants.

Rocks susceptible to the development of karst are to a greater or lesser extent soluble in water. In the UK four main types of karstic rocks – limestone (including chalk), dolomite, gypsum (including anhydrite) and salt (Cooper et al., 2011) occur. Soluble gypsum, anhydrite and salt, frequently referred to as evaporites because they were deposited under arid conditions by evaporation, are included here because they are subject to karst processes, though in the UK's relatively humid climate, these are rapid. Each presents a different erosion rate related to rock solubility and strength. As well as denudation, dissolution may result in the formation of cavities and cave systems and lead to subsidence of the surface. Active karsts only develop where the soluble rocks are exposed at the surface and/or where they are subject to active groundwater flow that results in dissolution. In the UK, in the case of carbonate rocks (limestones and dolomites) this is usually confined to the upper 100 to 150 m and is rarely reported from depths greater than 200 m even when the carbonate rocks are thicker than this. Palaeo-karsts (those formed in earlier geological times when the affected rocks were close to the surface at the time of formation) are widespread but are 'fossilised' and not active under current conditions. Evaporites are very much more soluble in fresh water than carbonates and dissolve rapidly when in contact with fresh surface or groundwater and as a result, in the UK, are not found outcropping at the surface. They usually occur in clay rich sedimentary successions and are protected from circulating groundwater by these rocks. Where dissolution occurs near to the surface, usually at depths of less than 100 m, subsidence is common.

Location	**Climate**	**Setting/geology**	**NDA rock Description**	**Authors**	**Method**	**Rate** m Myr^{1}
Dry Valleys, Antarctica	Polar	Crystalline	Higher Strength	Summerfield et al. 1999	^{21}Ne	0.26-1.02
Antarctica	Polar	Sandstone (hyper-arid)	Lower Strength	Nishiizumi et al.1991	^{10}Be and ^{26}Al	0.1-1.0
S. Norway	Sub-arctic	Elev. Plain gneiss mica schist	Higher Strength	Nicholson 2008	Quartz veins, weathering rinds, Schmidt hammer	0.5-2.2
Eyre Peninsula, Australia	Mediterranean	Granite (semi-arid)	Higher Strength	Bierman and Turner, 1995		0.5-1.0
Pajarito Plateau (NM)	Temperate	Tuff (temperate)	Higher Strength	Albrecht et al. 1993	^{10}Be and ^{26}Al	1-10
N. Sweden	Sub-arctic	Plain Crystalline	Higher Strength	Stroeven et al. 2002	^{10}Be and ^{26}Al	1.6
Canada	Polar	Plain crystalline rocks	Higher Strength	Peulvast et al. 2009	Palaeo-surface recon-struction	2-8
Masanutten Ttn, USA	Temperate	Sandstone, Shale	Lower strength	Afifi and Bricker, 1983	Mass Balance	2-10
S. Piedmont, USA	Temperate	Piedmont, granite	Higher Strength	Pavich, 1986	Mass Balance	4
Baltimore Piedmont, USA	Temperate	Piedmont granite	Higher Strength	Cleaves et al. 1970	Mass Balance	4-8
Brubaker Mts, USA	Sub-arctic	Low relief Schist, gneiss	Higher Strength	Price et al. 2008	Mass Balance	4.5-6.5
Rheinsh Massif, Germany	Temperate	Sedimentary	Lower strength	Meyer et al. 2008	^{10}Be	4.7-6.5
Iceland	Sub-arctic	Basalt	Higher Strength	Geirsdottir et al. 2007	Sediment record	5
Namib desert, S. Africa	Mediterranean	Granite Inselbergs	Higher Strength	Cockburn et al. 1999,	^{10}Be and ^{26}Al	5-16
Haleakala and Mauna Loa (HW)	Temperate	Basalt (various 0-3km elevation)	Higher Strength	Kurz, 1986		7-11
Mt Evans (CO)	Sub-Arctic	Granite erosion of bare surface (Alpine)	Higher Strength	Nishiizumi et al. 1993	^{10}Be and ^{26}Al	8
Georgia Piedmont (GA)	Temperate	Granite (temperate)	Higher Strength	Bierman, 1994	^{10}Be and ^{26}Al	8[1]
S. Piedmont, USA	Temperate	Piedmont granite	Higher Strength	Pavich, 1989	^{10}Be and Residence time	20
Luquillo Experimental	Temperate	Quartz diorite	Higher Strength	Brown et al. 1995	^{10}Be	25
Pacific NW, USA	Temperate	Orogenic	Higher Strength	Dethier, 1986	Mass Balance	33
S. Blue Ridge	Temperate	Schist, gneiss	Higher Strength	Velbel, 1985	Mass Balance	37
Smokey Mts, USA	Temperate	Schist, gneiss	Higher Strength	Velbel, 1986	Mass Balance	38

Boso Peninsula Japan*	Humid-temperate	Sandstone/ mudstones High rates of glacio/eustatic change	Lower Strength	Matsushi et al. 2006	^{10}Be, ^{26}Al	90-720
European Alps*	Sub-arctic	Orogenic	Higher Strength	Vernon et al. 2008	AFT	200-700
India*	Mediterranean	Escarpment	Higher strength	Gunnell, 1998	Functional Relationship model	205-275
San Bernadino Mountains, California, USA*	Temperate	Orogenic, quartz-monzonite and granodiorite, sandstone, granite	Mixture	Binnie et al. 2008	^{10}Be, Apatite (U-Th/He) thermochrono-metry	70-1200

Table 6: Denudation rates of bedrocks from climates around the world (m Myr^{-1}).
* Note that these are areas of active mountain building which consequently have much higher denudation rates than occur in mid-crustal locations such as the UK.

When considering bedrock lowering rates it is important to remember that most bedrock is covered by a mantle of soil. Bedrock lowering rates under soil have also been examined using cosmogenic radionuclides and have been used in describing the soil production function, which describes the rate of soil production (or bedrock lowering) as soil depth increases. Ranges of soil production vary greatly in individual studies (Table 7). A major reason is a result of a negative feedback loop that occurs in the weathering of bedrock under soil. For example, Heimsath *et al*. (1997) have found that if soil thickness develops beyond a certain depth, then moisture can no longer reach the bedrock to maintain the weathering reactions. Conversely, if the soil is too thin then the moisture rapidly evaporates. Riggins et al (2011), in a compilation of results, suggests greatest production is found when soil depths are between 15 and 30 cm deep. At 40-50 cm depth soil production rates decrease. Rates of bedrock lowering via the soil production function are similar to non-orogenic bedrock denudation rates (generally < 50 m Myr^{-1}) and this is probably because of the fact that precipitation will run off bare bedrock. However, in the context of a GDF, both bedrock lowering and soil production rates give a reasonable estimate of the denudation rates of the Earth's surface.

Location	Author	Elevation (m)	Rock type	NDA rock description	Mean Annual Temperature	Mean annual precipitation (mm)	Regolith Production rate (m Myr^{-1})
Arnhem Land (Australia)	Heimsath et al. (2009)	150	Sandstone	Lower strength	28.5	1400	6.3-25.8
Blue Mountains (Australia)	Wilkinson (2005)	1000	Sandstone	Lower strength	10.5	950	9.1-18.2
Bega Valley (Australia)	Heimsath et al. (2000)	400	Granite	Higher Strength	20	910	9.3-68
Cornwall (England)	Riggins et al. (2011)	370	Granite	Higher Strength	10	1250	9.6-20
Point Reyes (California)	Heimsath et al. (2005)	300	Granodiorite	Higher Strength	12	800	11-110
Wind River Range (Wyoming)	Small et al. (1999)	3600	Granite	Higher Strength	0.5	650	12.6-14.8
Coast Range (Oregon)	Heimsath et al. (2001)	250	Sandstone	Lower Strength	10	2000	14.7-359
Tennessee Valley (California)	Heimsath et al. (1997)	120	Greywacke	Higher Strength	15	760	15-107

Table 7 Rates of soil production or bedrock lowering under a soil mantle

5.5.2 Estimates of denudation from catchment studies

Estimates of surface denudation have also been determined by monitoring the outputs of catchments. Catchment outputs can be divided into three types; i) chemical solute flux; (ii) suspended sediment and (iii) bedload. Estimates of denudation are often based on the area of the catchment and presented as t km^{-2} yr^{-1} which can be used to calculate an average denudation rate over the catchment under consideration. Comparison of the denudation rates determined from catchment studies, by using cosmogenic radionuclide methods and from soil production rates shows that the different approaches provide similar results.

5.5.2.1 SOLUTE FLUXES

Landscape denudation within catchments may be measured using solute fluxes in stream water close to the exit point for a stream from a catchment. Table 8 lists values for solute fluxes from catchments in a range of climates. By converting these factors, assuming a bulk density (g cm^{-3}) of one, denudation from measured solute fluxes range from 1 to 122 m Myr^{-1} with values for arctic environments being lowest and those for karst areas in tropical/monsoon environments being highest.

Location	Climate	Geology	NDA rock description	Solute Flux (t km^{-1} yr^{-1})	Authors
Canadian Shield and interior Basin	Arctic / Sub-arctic	Sedimentary / Volcanic	Lower – Higher Strength	1.0 – 5.3 (very low)	Rosa et al. 2012
Strengbach catchment, Vosges Massif, France	Temperate (oceanic-mountainous)	Ca-poor granite	Higher Strength	2-4.7.6 (estimates)	Viville et al. 2012
Kidisjoki catchment, Finland	Arctic	Gneisses and granulites	Higher strength	2.9	Beylich et al. 2006
South Cascade Glacier catchment, U.S.	Alpine – 32% covered in glacier, 1992	Migmatite, diorite, gneiss, schist, amphibolite	Higher strength	Cation flux = 14.1 SiO_2 flux = 4.0	Axtmann and Stallard (1995)
Haut Glacier d'Arolla, Switzerland	Alpine – 54% catchment covered in glacier, 1989			Cation flux =12.9; SiO_2 flux = 4.2	Sharp et al., 1995
Narmada River Catchment, India	Humid Tropical	Basalt + minor carbonates and saline – alkaline soils)	Lower – higher Strength	25-63 t km-2 yr-1	Gupta et al. 2011
Guizhou Province	Mild-humid monsoon	Karst	Lower strength	56 t km-2 a-1	Yongbin and Hongbing, 2011
Jiangxi Province, China	Sub- tropical monsoon	Karst	Lower strength	122 t km-2 a-1	Yongbin and Hongbing, 2011

Table 8: Solute fluxes from a range of climates and geology types

5.5.2.2 SUSPENDED SEDIMENT YIELDS

Denudation can also be measured in studies through suspended sediment yields, normally measured at catchment exits. In the last 20 to 30 years, many soil erosion rates on grassland and arable sites in the UK have been reliably estimated using the redistribution of ^{137}Cs released from nuclear weapon testing. Walling and Webb (1987) suggest that the current range of suspended sediment yields in the UK varied between <1.0 and 488 t km^{-2} yr^{-1} with an average of about 50 t km^{-2} yr^{-1}. This equates to about 50m using the assumptions above. Extrapolating these data suggests that erosion will be higher in periglacial conditions when there is lower vegetation density and higher power streams.

5.5.2.3 RIVER EROSION / INCISION

River incision during the Quaternary (last 2.6 million years) can be determined from the height of river terraces above the current river bed. Rivers in Southern England and Northern France, where the land is currently subsiding in the south and rising in the north show that maximum depth of incision is 165 m within this time frame (2.6 Myr). Table 9 summarises information for major rivers in Southern England and France where the land is subsiding. Data are taken from Bridgland (2010).

River system	Number of terraces	Current height between top terrace and river
Fenland Rivers	5	23m
Lower Thames	4	38m
Trent	6	65m
Somme	12	70m
Seine	6	70m
Solent	22	80m
Severn-Avon	8	80m
Upper Thames – Evenlode	8	135m
Middle Thames	14	165m

Table 9: Difference in height in the lowest to highest river terraces from rivers in Southern England and Northern France. Data collated from Bridgland (2010).

5.5.3 Erosion by river capture

River capture is a geomorphological process where an existing river system is diverted and a new drainage route established through merging with a neighbouring river as a result of (i) tectonic activity, (ii) natural damming such as by ice sheets, (iii) headward erosion of one stream into another, (iv) lateral erosion by meander or (v) within karst landscapes where underground dissolution may re-route rivers. River capture can therefore increase the catchment size of the active 'capturing river', increasing the power, flood magnitude and frequency. Thus river capture can potentially increase flow and therefore erosion rates. Examples of river capture in the UK include the river Thames which underwent a series of changes through the Pleistocene (Bridgland, 1994), the River Stiffkey in Norfolk (Brand et al. 2002), the River Teme (Cross and Hodgson, 1975) and the River Trent.

The potential to affect a GDF as a result of river systems (i) changing course and incising into areas where GDF infrastructure may exist and (ii) through the potential changes in groundwater dynamics because of the alteration of a river's course will be limited to the near surface parts of its infrastructure. Maximum incision as a result of these processes is likely to be similar to that as a result of incision discussed above, around 30 m over one million years. The maximum incision (middle Thames) over the Quaternary per million years is about 63.5 m.

5.5.4 Glacial erosion – Landscape analysis

5.5.4.1 Glacial Troughs, fjords and overdeepening

In upland areas, ice streams can erode channels with parabolic cross sections (troughs and fjords) up to depths of 2 km (Talbot, 1999) and thus they need to be considered in terms of their effect on a GDF. The depth of the trough or fjord represents the effect on ice erosion rate of the positive feedback between pressure gradients induced by the topographies of the top and bottom of the ice and the temperature and velocity of basal ice. Glacial overdeepening is a characteristic of valleys previously occupied by glaciers and can be hundreds of metres deeper than the thalweg (deepest continuous line) along a valley or watercourse. Glacial overdeepening can create valleys deep enough to potentially affect a GDF (if sited at the shallow end of the proposed depth range of 200 to 1000 m) or the backfilled infrastructure. For example, Hall and Glasser (2003) examined Glen Avon, a 200 m deep glacial trough in Scotland.

	Volume (km^3)	**Area (km^2)**	**Length (km)**	**Max depth (m)**	**Mean depth (m)**	**Dominant Bedrock Geology**
Loch Shin	0.35	22.50	27.8	49	15.5	Psammite
Loch Awe	1.2	39	41	94	32	Quartz-arenite
Loch Lomond	2.6	71	36	190	37	Psammite and pelite
Loch Maree	1.09	28.6	20	114	38	Sandstone and mudstone/ mafic lava and fuff/Schist
Loch Shiel	0.79	19.5	28	128	40	Semi-pelite and pelite
Loch katine	0.77	12.4	12.9	151	43.4	Psammite and Pelite
Loch Arkraig	0.75	16	19.3	109	46.5	Psammite/ felsic rock
Loch Rannoch	0.97	19	15.7	134	51	Psammite and semi-pelite
Loch Ericht	1.08	18.6	23	156	57.6	Psammite and semi-pelite, pelite
Loch Tay	1.6	26.4	23	150	60.6	Psammite
Loch Lochy	1.07	16	16	162	70	Psammite and pelite, psammite and semi-pelite, Old Red Sandstone
Wastwater, Lake District				79		
Loch Ness	7.45	56	39	230	132	Psammite/ sandstone/conglomerate/

Table 10: Dimensions of Scottish Lochs and Lake District lakes

Fjords are typically flooded overdeepened glacial troughs. Overdeepening develops by the quarrying, plucking, abrasion and glacial meltwater erosion. The overdeepening process is considered to exert major controls on the glacial landscape. The importance of overdeepened troughs is that they are likely to influence the response of future ice masses. In addition, overdeepening may increase with successive glacial periods. This process has been responsible for forming, for example, the Scottish lochs and the fjords of Scandinavia. The depths of the Norwegian fjord Geirangerfjord descends to over 600 m below sea-level and Aurlandsfjord, also in Norway, is up to 1308 m deep and is considered the deepest in Europe. This amount of overdeepening is considerably more than is seen in the UK; some example figures are given in Table 10, the maximum of which is the 230 m of Loch Ness which is the result of multiple glaciations. Such overdeepening can be enhanced if it occurs close to faults as in the case of Loch Ness, which is developed on the Great Glen Fault.

In lowland areas overdeepening of valleys can be attributed to:

- Steady state drainage of meltwater and groundwater driven by hydrostatic pressure gradients;
- Catastrophic outbursts of water (jokulhlaups); and
- Glacial erosion. Overdeepened valleys can sometimes be buried or submerged.

In Norfolk, buried valleys can reach a depth of 100 m and it has been suggested that they were formed by sub-glacial streams incising into the chalk (Woodland, 1970). Huuse and Lykke-Anderson (2000) reported on a series of offshore overdeepened and buried quaternary valleys in the eastern Danish North Sea. These were identified by high resolution multichannel reflection seismic survey and some borehole data. It is believed that these deeply incised valleys had their origin in repeated cycles of sub-glacial down-cutting of meltwater channels and channel closure as a result of ice creep. The authors suggest that a major requirement for their formation is a poorly consolidated substrate whilst changes in permeability can produce sub-glacial variations in their depth. The diagnostic size features of the Quaternary valleys in the Danish North sea include (i) they are all buried with sediment, (ii) valleys can be incised to a depth of 300 m below present day sea-level, (iii) valleys are typically 1-5 km wide and the general depth to width ratio is ~1:10.

Further important discussion was made by Huuse and Lykke-Anderson (2000) on the distribution of buried Quaternary valleys in northwest Europe. Firstly, they suggested that eskers and deep valleys rarely co-existed in the same area. For example, valleys (except fjords) are rare outside the area covered by poorly consolidated Mesozoic and Cenozoic sediments. However, eskers are most common in areas of lithified sediment or crystalline rocks, suggesting that valleys and eskers are both formed by sub-glacial water but their different geomorphological forms are produced on substrates of different erodibility. The permeability of the Cenozoic sediments may have also contributed to the character of sub-glacial erosion and valley distribution. For example, the Oligocene sediments are coarse-grained and are relatively permeable in comparison with the Miocene sediments. Thus, greater sub-glacial drainage would have occurred as groundwater in the Oligocene sediments, thus reducing down-cutting and producing shallower valleys.

Fichler et al. (2005) used seismic and magnetic data to locate sub-glacial meltwater channels of various dimensions in Quaternary strata in the Norwegian Central North Sea. In addition crater-shaped depressions with diameters ranging from 500-3000 m and with depth of 20-300 m were found coinciding with the initiation of the channels. They suggested that these craters were generated by gas expulsion from melted gas hydrates combined with melt-water expulsion and erosion. Evidence supporting this suggestion was that:

- The craters are similar to ones recognised as being formed by methane hydrate release;
- Appropriate conditions existed for formation and melting of gas hydrates during glacial and de-glacial periods; and
- Correlation with shallow gas occurrences and seismic gas indications.

The existence of buried valleys both on- and off-shore demonstrate the depths to which glacial meltwater erosion can occur. In some circumstances the depth of these valleys is within the minimum depth (200 m) for a GDF. However, the high rates of erosion are likely to be associated with existing glaciated valleys; these areas can be avoided for the siting of a GDF

5.5.4.2 ICE SHEET EROSION

Ice sheets flex the lithosphere and lead to fluctuations in patterns of stress and permeability as well as sea-level. They have the potential to affect a GDF because they are capable of removing significant thicknesses of material from the surface of the Earth, depending on ice thickness, geology and length of time the ice sheet is dynamic. In the Pleistocene, large ice sheets could exceed 3 km in thickness and a diameter approaching 1000 km in (as in current Greenland) and 4000 km in Laurentia (as in current Antarctica). Smaller ice sheets were found in Iceland and Scotland and the mountain chains of Europe. The thickest ice had the potential to depress rockhead up to about 960 m but, because the resulting flow in the asthenosphere is slow, it is likely that equilibrium was never reached, and the maximum depression attained was about 600 m (Boulton and Payn, 1993).

Several studies have examined ocean deltaic deposits to assess the degree of terrestrial erosion during glacial periods. Dowedeswell *et al*. (2010) examined the 2.7 Myr old Naust formation situated offshore of mid-Norway using dating and siesmic surveys. The 100 000 km^3 deposit is considered to have been mostly deposited as a result of glaciations. For example the part of the sequence covering the period 0-0.2 Myr BP (including the Weichselian and Saalian glaciations) is equivalent to 16,300 km^3 of material. Calculations suggest that this is equivalent to a bedrock erosion rate of 0.41 m $k.yr^{-1}$ or 81 m of bedrock erosion across the ice sheet catchment. Across the whole 2.7 Myr period, bedrock lowering is estimated to be 524 m with an average bedrock lowering rate of 0.19m k yr^{-1}. The results also suggest that the hypothesis that greatest erosion occurs at the beginning of an ice age when most rock is available for erosion is not true in this case. They suggested that the mean sediment delivery was two to three times greater for the most recent 600,000 years than for the earlier parts of the Naust formation.

Thus changing ice sheet size and dynamics and the intensity of individual glacial cycles appear to be important. The authors compared the long term accumulation rates from the Naust formation with other major marine sediment depo-centers such as those from the Amazon and Mississippi rivers since the late Cenozoic. They found that that the Amazon (0.01 m kyr^{-1}) and Mississippi (0.025 m k yr^{-1}) were an order of magnitude lower than those from the Barents Sea ice sheet (0.24 m kyr^{-1}) and the mid Norway ice sheet (0.24 m kyr^{-1}). Clayton (1996) reviewed the off-shore sediments of the UK and estimated an average depth of erosion from present land surface and the adjacent continental shelf during the Quaternary. For the northern UK, which has been repeatedly glaciated, an estimate of eroded thickness lies between 125 and 155 m, the latter figure representing the erosion of softer sedimentary rocks. Estimates from this study suggest that over the last 440,000 years, ice has eroded the land surface at least twice as fast as rain and rivers and could be up to five fold faster.

Other weathering/erosion processes need to be considered with respect to glaciers and ice sheets and are particularly relevant to new groundwater movements which can produce new rock-water interactions. Both changing groundwater pathways and new rock-water interactions caused by weathering processes could have effects on a GDF infrastructure.

Ice pressure from ice sheets or glaciers can cause 'hydraulic jacking'. This is caused when water pore pressure in a fracture exceeds the normal stress acting on it and the fracture's tensile strength. This can cause the fracture to dilate and propagate (Lönnqvist and Hökmark (2010). The fracture surfaces can separate and the openings can become large, therefore affecting both flow and transport processes in the rock. Lönnqvist and Hökmark (2010) modelled three scenarios for SKB, these being:

- Advancing or stationary ice front without permafrost;
- Advancing ice front in combination with proglacial permafrost; and,
- Retreating ice front without permafrost.

In the second, most conservative, scenario they suggest a maximum jacking of 350 m. They suggest that the potential for hydraulic jacking is always greatest at the surface and is likely to be initiated in the ground surface. The process can be reversible as long as there is no shear component or the fracture is filled with sediment (Hökmark et al., 2006). Glacial hydraulic jacking occurs at the initiation of the ice sheet when high pore pressures build up in the rock underneath the ice sheet. As the ice sheet advances, high pore pressures can be transferred to the ice front by long highly transmissive fractures or they can build up below an impermeable permafrost layer in front of the ice sheet.

During deglaciation some of the glacially induced pore pressure can be retained by the rock. The extent of glacial lift can be considerable. Pusch et al. (1990) modelled hydraulic uplift and suggested that subglacial artesian water flowing 1 km behind a 200 m high ice front could potentially lift already loosened rock slabs 30 m thick (between 14 and 33% of the ice thickness). Talbot (1999) points out that the Weischselian ice sheet was close to 3,500 m thick over much of Scandinavia, thus potentially causing hydraulic jacking of rock slabs 500-1000 m thick to be lifted. In addition, Talbot (1999) also describes the infilling of sub-horizontal fracture zones with clay rich materials found in boreholes of 1000 m depth from Fennoscandia. The source of this fracture filling material remains to be determined but is considered to be either fault gouge or superficial Pleistocene sediments injected deep into the fracture zones by overpressured glacial meltwater.

Fracturing of bedrock by hydraulic pressure is likely to control groundwater flow in any rock of low porosity that deforms in a brittle manner, such as igneous and metamorphic rocks and sedimentary rocks that have been cemented or are well consolidated. Hydrofracture systems represent a visible expression of the passage of pressurized meltwater through sub-glacial bedrock to ice-marginal environments (Vaughan-Hirsch et al. 2012). Hydrofractures represent the fluctuations in hydrostatic pressure within the glacial hydrogeological system in the host sediment/bedrock. Increased pressure within this system can result in accelerating rock fracture by the redistribution of stress on sub-glacial rock and changing the pressure of water in cracks (Iverson, 1991). Kamb et al. (1985) measured basal water pressures in boreholes and they have been found to diurnally fluctuate by more than 0.5MPa. In addition, the pressure of sub-glacial waters influences the sliding velocity of glaciers and quarrying of subglacial bedrock (Iverson, 1991). Hydrofracturing may also facilitate the initial detachment and transport of sediment or bedrock rafts, including megablocks. These are large blocks of sediment or bedrock which have been dislocated and transported a considerable distance from their source outcrop by glacier ice. Examples in the UK are the chalk glaciotectonic rafts in Norfolk (Vaughan-Hirsch, 2012; Burke et al. 2009). These can be up to 80 m long and 15 m thick. Chalk and limestone are particularly susceptible to lifting because of their generally blocky nature.

5.5.5 Permafrost and periglacial weathering and erosion

Permafrost is defined as ground that is ‘permanently frozen, or that remains below freezing temperature for two or more years’ (as defined by the International Permafrost Association at http://ipa.arcticportal.org/resources/what-is-permafrost). The physical description of frozen ground is traditionally based on a two-layer conceptual model based on a seasonally frozen ‘active layer’ overlying perennially frozen materials. Shur *et al.* (2005) suggest that this model is inadequate to explain the behaviour of the active zone/ permafrost over long periods, particularly in ice-rich terrain. Thus a third term, the ‘transition zone’ is used to describe the zone that alternates in status between seasonally frozen ground and permafrost over a sub-decadal to centennial time scale. The transition zone can be described as being ‘ice enriched, and functions as a buffer between the active layer and long term permafrost by increasing the latent heat required for thaw’ (Shur et al. 2005). Thus it imparts stability to the permafrost under low amplitude or random climatic fluctuations.

5.5.5.1 Physical properties of the active layer and weathering processes

Active layer development is the seasonal depth to which thawing occurs in cryogenic environments. Tye et al. (2012) suggest that it was the annual formation of ice within the active layer that was a major factor in controlling soil depth above the Sherwood Sandstone outcrop in Nottinghamshire and the active layer development was considered to extend to depths about 16 m. A GDF at greater than 200 m depth will not be affected by active layer processes.

An important aspect of weathering and erosion in areas subject to periglacial conditions is the process of freeze thaw or frost shattering. This process will be one of the factors that enhances denudation under cold climatic conditions.

5.5.5.2 Physical properties of permafrost

Permafrost processes in the ‘active layer’ are well understood but less is known about weathering processes in deeper (100s m) permafrost. Table 11 provides data on the depth to which permafrost has been found (UK permafrost depths are expected to be at the shallower end of the depths given in this table). Because permafrost may reach depths of several hundred metres in the UK it may affect a GDF. However, because permafrost develops from the surface the expansion of water on freezing will be accommodated by flow through unfrozen pores. This may induce slightly enhanced groundwater flows through adjacent unfrozen rock. In the UK geology can play a part in the formation of permafrost. For example, Hutchinson and Thomas-Betts (1990) examined the extent of permafrost in southern England at approx. 18,000 years BP. It was noticed that permafrost features were largely absent from large parts of SW England and this is attributed to the high heat flows produced by the granite batholiths.

5.5.5.3 Potential weathering and erosion changes that could affect GDFs related to permafrost.

There is a paucity of information on bedrock weathering processes through permafrost interactions at deep (100’s of metres) depth which may impact on a GDF. However, an understanding of permafrost is particularly important because it is likely that the depths to which it can develop will be within the range of a GDF infrastructure (200 to 1000 m). In particular, if the buffer or backfill were to freeze under permafrost conditions, this would imply that there is a mechanical load on the surrounding rock (SKB, 2006b). McEwen (1991) suggested that permafrost rarely exceeds 500 m so that a GDF is unlikely to freeze in an environment with a thermal regime of 30°C km^{-1}. The average UK geothermal gradient is 26°C km^{-1}, but locally it can exceed 35°C km^{-1} (Busby; 2010). However, GDF features within the permafrost zone (e.g. backfilled tunnels, shafts) and associated materials (e.g. cements), and within the freezing zone could be effected. McEwen (1991) assumes the concrete would become badly degraded by freezing cycles whilst the effects of permafrost on bentonite clay are harder to predict. The following effects may occur:

- Thermo-osmosis will move water out of the clay according to the thermal gradient;

- The decrease in temperature towards 0°C will increase viscosity of the water in the clay's pores and will tend to decrease its hydraulic conductivity;
- A lowering of the temperature will decrease the clay's pore size and will lead to water movement away from the clay as the freezing front approaches;
- Thermal contraction of the clay will take place; and
- The clay will become less plastic as temperature decreases.

However, of more concern may be the impact on any bentonite present in a GDF. An increase in brittleness in the bentonite, combined with its contraction may reduce its effectiveness as a seal. During deglaciation, rapid ice removal may result in fault movement and dilation of fractures. In the UK such fault movement will be small (see Chapter 3) and in the unlikely event of a fault that intersects a GDF being reactivated the bentonite backfill is likely to be able to accommodate the displacement.

Location	**Type of Permafrost**	**Depth of Permafrost (m)**	**Source**
Finland (29 Locations)		Max 100, 10-50 mainly	King and Seppälä, 1987
Svalbard Coastal regions		<100	Humlum et al. 2003
Svalbard		220	Isaksen et al. 2001
Southern Sweden		350	Isaksen et al. 2001
Svalbard Highlands		500	Humlum et al. 2003
Lupin Mine, Nunavut, Canada	Basal	550-570	Stotler et al. 2009 a, b
High Arctic islands, Canada		600	Collet and Bird, 1988
Northern Siberia		900	Schwartsev et al, 1988

Table 11: Review of permafrost depths in a variety of alpine and polar locations.

The impact of permafrost on mechanical processes

Ice growth in bedrock could result in new groundwater pathways being created by opening or closing joints, fissures and cracks. Small amounts of evidence regarding ice formation in deep bedrock fractures (100's of metres) have been obtained via mining investigations in permafrost areas. Ice in bedrock has been observed at the Asbestos Hill mine, northern Quebec where fractures in the rock were found to be ice filled with up to 50 mm of ice (Samson and Torden, 1969). Zhukovskiy et al. (1973) described ice in cracks of granite. They suggested that expansion of the cracks occurred during freezing and thawing of the rock. It is the expansion of small cracks and fractures that could provide new pathways for water and gases causing weathering and precipitation reactions (see below).

The impact of permafrost on chemical processes

Changes in climate or the mechanical processes described above could result in re-routing of groundwater, thus affecting groundwater stability, which is probably the principal driver of sub-surface erosion and weathering processes (precipitation and solubility of minerals). These could include:

- Changes in groundwater circulation resulting from density changes as permafrost melts and interacts with brines and cryopegs (saline waters);
- The formation of cryopegs through the loss of water to the ice front, thus leaving the residue waters as being very saline;
- The melting of permafrost by brines and cryopegs;
- The melting of basal permafrost increasing the inflow of fresh and saline waters into circulation;
- Increased mixing of fresh – saline – brine waters causing precipitation and solubility of minerals in cracks and pores;
- Possible saline intrusion in coastal regions resulting in changed groundwater circulation, including as a result of tectonic changes leading to weathering; and
- Changing redox status of groundwater as a consequence of increased mixing or recharge events leading to weathering.

When water freezes to form ice new minerals are often formed. Gravitational, capillary and loosely bound non-saline water freezes at ~0°C, whereas bound water can freeze within a range of negative temperatures. Salt water with total dissolved solids (TDS) 30 g l^{-1}, crystallize at about -1.5 to -2°C and brines can remain liquid at -20°C or lower. The freezing of water causes a differentiation of salts between the solid and liquid phases. Fractions of the salts are (i) contained in the ice, (ii) the least soluble salts which precipitate and (iii) the easily soluble salts which are squeezed into lower water layers, thus increasing the TDS. Ice formed by freezing has significantly lower TDS concentrations than the initial pore solution. Taking into account the degree of solubility at negative temperatures $CaCO_3$ would be expected to precipitate first (temp range -1.5 to -3.5°C), and then $NaSO_4$ and $CaSO_4$ (temp range -7 to -15°C). Consequently, cryogenic layers can be enriched with gypsum, mirabilite and calcite. Below the freezing layer cryopegs can form through the exclusion of solutes downwards.

The potential effects of these water-rock interactions are examined above, but the fundamental effect is that many of these reactions could alter water and gas pathways that allow transport to the surface and towards a GDF barrier.

5.6 BIOSPHERE EFFECTS

5.6.1 Effects of biosphere on terrestrial weathering and erosion

Vegetation and microbial activity enhances weathering in terrestrial systems through (i) the production of organic acids, (ii) changing redox status and (iii) physical and chemical weathering by roots. The degree to which these contribute to the weathering of minerals under current climates is part of the solute weathering load referred to previously, but the extent to which these processes contribute to this total load is unknown. Analysis of erosion rates taken in UK current climates demonstrates the role of vegetation in preventing erosion. Net grassland soil erosion rates measured are reported to be in the region of <1 t ha^{-1} (DEFRA, 2006) compared to those from arable fields which range up to 15 t ha^{-1} dependent on hill slope, hill length and length of time the soil is bare (DEFRA, 2006; Quine and Walling, 1991). Vegetation acts to hold the mobile regolith/soil together through their roots and also intercept rainfall, thus reducing energy of raindrop impact. Vegetation also plays a role in reducing pore pressure through the evapotranspiration cycle, thus reducing moisture content in soils which can help prevent landslides. Thus, based on the role of vegetation in regulating erosion, the greatest differences in erosion rates are probably found between landscapes which are vegetated and those which are not. Thus in polar landscapes where there is abundant material to be eroded and often little vegetation cover, erosion rates would be considered higher.

5.6.2 Effects of deep subsurface biosphere on weathering and erosion

The Biosphere Status Report (NDA, 2010b) primarily considers the potential releases of radionuclides from a GDF that may reach the surface biosphere by transport in groundwater and gaseous phases. Issues relating to microbial gas production, radionuclide migration and behaviour in the biosphere (atmosphere, surface and near-surface environments) are discussed, however, there is no significant reference to processes occurring in the deep subsurface biosphere. This section will summarise the impact of natural processes on the deep biosphere and how these may affect microbial activity, which may ultimately impact on the long-term performance of a GDF.

This summary will:

- Briefly summarise the geomicrobiology of relevant geological environments;
- Identify the microbial metabolic processes that are likely to be prevalent in natural host rock formations representative of the far-field and how these processes impact on radionuclide mobility.
- Establish how perturbations in the deep subsurface resulting from natural processes may affect microbial processes and consequently impact on radionuclide mobility following disposal of radioactive waste.

Natural geological processes that may impact on a GDF resulting in physical and chemical perturbations which have been described in other chapters include:

- Tectonic movements;
- Uplift (rebound), or subsistence;
- Weathering and erosion; and
- Climate change especially future glaciations.

These processes have the potential to induce biogeochemical and physical changes through the introduction of new chemical species, organic matter, gases and surface (exogenous) microorganisms into the deep subsurface.

5.6.3 Microbial ecology in the deep subsurface

It is well recognised that microbes live in a wide range of subsurface environments, even if growth is strongly constrained by limited nutrient and energy supplies resulting in very low metabolic rates (D'Hondt et al, 2002; Lin et al, 2006; Roussel et al, 2008, West and Chilton, 1997). Indeed, it has been estimated that the mass of subsurface microbes may exceed the mass of biota on the Earth's surface (Whitman et al, 2001).

Consequently, characterisation of many geological environments relevant to the geological disposal of radioactive waste have been undertaken. The UK radioactive waste microbiology programme is one of the longest running, having started in the early 1980s, and pioneered 'radioactive waste geomicrobiology'. Surveys of relevant geological environments have been an important component of the work in the UK and have involved studies in continental Europe and in Britain. Granites (UK, Sweden) were also assessed together with the Boom Clay (Belgium), Salt (Asse, Germany) and sedimentary sites (UK and Europe) (see Humphreys et al, 2010 and references therein). Other national programmes have also undertaken characterisation work in relevant geological formations: granites (Canada, Finland, Japan, Sweden, Switzerland); sedimentary rocks (Belgium, Germany, Hungary, Italy, Japan Switzerland); evaporates (Germany, Switzerland); volcanic tuff (USA); and salt formations (USA). More detailed information can be found in Humphreys et al (2010) and in the references therein. Analyses have mostly focused on groundwaters, although solid materials have also been investigated for microbial content. In brief, low populations of microbes in complex ecosystems were demonstrated in every environment (e.g. Ahonen *et al*., 2010; Hallbeck and Pedersen, 2008; Hallbeck and Pedersen, 2008; Itavaara *et al*., 2011; West and McKinley, 2002; Pedersen, 2000; Wersin *et al*., 2011). The exact composition of the ecosystem is site specific, but these studies do show the potential range of organisms likely to be detected in a particular geological environment. Functional groups include sulphate reducing bacteria, nitrate reducers, iron reducers, acetogens, methanogens, halophilic groups; viruses (Humphreys et al, 2010).

The environmental controls of all microbial populations in any UK geological environment will be:

- The availability of nutrient and energy sources for microbial usage;
- Groundwater flux;
- The geological history of the site, including recent usage (e.g. water abstraction, proximity to contamination sites);
- The specific geological environment, e.g. geochemistry.

Thus any natural changes impacting on these environmental controls will alter the subsurface ecosystem in any given environment.

5.6.4 Microbial metabolism in the deep subsurface

A GDF will inevitably be placed within a subsurface ecosystem and there will be complex biogeochemical interactions in the near-field (West et al, 2002), particularly within the excavation disturbed/damaged zone (EDZ). It is also well recognised that microbes living in deep geological environments can also impact on solute transport processes in the far-field (Humphreys et al, 2010) and will thus be subject to influence by any natural changes. This work is the subject of on-going research (see summary in Humphreys et al, 2010). In the UK currently, work on microbial interactions in the EDZ and far-field is being funded through the Natural Environment Research Council as part of the Biogeochemical Gradients and Radionuclide Transport (BigRAD) and the new Radioactivity and the Environment (RATE) Programmes in association with the NDA and Environment Agency and the Science and Technology Funding Council.

Microbial activity in any environment, including the deep subsurface, is generally located on chemical or physical interfaces within biofilms. The impacts can be both physical (e.g. altering porosity) and/or chemical (e.g. changing pH, redox conditions). These impacts may result in intracellular or extracellular mineral formation or degradation (Coombs et al, 2009). Considerable work on the effects of biofilms on overall transport processes in a variety of relevant geological environments has now been undertaken including granites, mudstones and sandstones (e.g. Anderson et al, 2011; Tuck et al, 2006; Harrison et al, 2011; Wragg et al, 2012). The work indicates that changes in local hydrological regimes can be changed by the development of biofilms, particularly when additional nutrients are introduced, as may occur when natural changes occur.

The above processes could also influence transport of radionuclides in the far-field of a repository. The presence and metabolic activity of micro-organisms has been recognised as an important component for controlling radionuclide solubility/mobility (either directly or indirectly) in the deep subsurface. Processes include adsorption/precipitation, complexation/chelation, dissolution, oxidation/reduction reactions, and colloidal aggregation (Anderson *et al*., 2011; Bass *et al*., 2002; Behrends *et al*., 2012; Humphreys *et al*., 2010; Pedersen, 2000; Pedersen 2005; Sherwood Lollar, 2011; West and McKinley, 2002; West et al, 2002). In addition, increased microbial gas production e.g. carbon dioxide, hydrogen and methane could affect the physical performance of a GDF through pressure build up and enhanced transport of radionuclides (e.g. Humphreys et al, 2010; Stroes-Gascoyne *et al*., 2007).

A summary of the microbial processes that lead either directly or indirectly to altered radionuclide solubility and sorption (and hence mobility) include:

- *Changes of redox conditions* - Microbial processes can change the oxidation state of certain radionuclides through enzymically catalysed redox transformations. These changes can affect radionuclide solubility and sorption and therefore mobility. This is a key parameter, particularly for redox-sensitive radionuclides such as, uranium, plutonium, neptunium, selenium and technetium, where higher oxidation states generally have increased solubility.
- *Modification of pH conditions* - Under anaerobic conditions, fermentation processes can produce short-chain fatty acids, including acetic acid. These weak organic acids would have limited effect on pH, and are generally metabolised quickly. Microorganisms also metabolise nitrogen and sulphur to produce inorganic acids. The acidification effect results from the dissolution of nitrogen dioxide which leads to the production of nitric and nitrous acids. The reduction of dissolved inorganic acids by methanogens has also been reported to increase the pH of the localised environment. Overall, reduction in pH due to microbial activity is likely to be limited by the slow rates of microbial activity in the deep subsurface and by the buffering capacity of the host rock.
- *Changes in host rock mineralogy (formation or dissolution of minerals which act as sorbents)* - Formation of clay minerals can reduce the permeability of igneous rocks and retard the migration of radionuclides. Microbes can also affect the physical and chemical condition of the host rock e.g. colonisation by bacteria, which could significantly enhance weathering/dissolution of the host rock. As a result of microbial weathering, cracks and fractures may be enlarged, which allows increased transport of liquids and gases.
- *Production or consumption of radionuclide binding ligands* – Production of organic ligands and complexing agents by microorganisms can increase radionuclide solubility and mobility. In contrast, precipitation of dissolved radionuclides can also be caused by the microbial production of ligands. It is noted that the degradation of soluble complexes can lead to adsorption or precipitation of the released radionuclide. Ligand production can also promote biosorption and/or bioaccumulation, and the dissolution of mineral phases resulting in desorption.
- *Production of biocolloids or biofilms* – The presence of planktonic microbes in groundwater which act as biocolloids can increase radionuclide migration through sorption onto microbial surfaces. Biofilm adhesion to mineral surfaces can reduce radionuclide sorption leading to increased radionuclide mobility.
- *Microbial gas production* - Microbial communities can consume abiotic gases produced in geological environments (e.g. hydrogen utilization by sulphate reducing bacteria and methanogens). They can also be responsible for the production of biochemical gases that contribute to the total gas load (e.g. methanogenesis) in these locations.

It is difficult to evaluate the scale and significance of these processes. However, it could be envisaged that should natural changes be rapid in time or scale (e.g. onset of glaciation) causing a significant change in the hydrological regime then the potential introduction of nutrients and energy sources from the surface would significantly increase microbial metabolic activity.

5.6.5 The impact of microbial processes on radionuclide mobility

Natural geological processes, such as ice melt, will impact on microbial metabolic activity in host rocks representative of the far-field of a GDF by causing changes in the supply of nutrients, water and electron donors and acceptors (particularly oxygen, hydrogen and carbon dioxide) (Bass et al 2002). The particular effect of oxygen in groundwater in stimulating microbial corrosion of waste containers could also be considered if hydrological regimes are significantly altered over a rapid timescale. However, it is likely that corrosion of waste canisters will occur soon after repository closure when oxygen is still available so this process is not considered further in this section.

Any natural geological process that can change the supply of nutrients, energy and water will alter microbial activity (and population density) and may, as a result, change the permeability of the system, alter radionuclide solubility and sorption processes and formation of biocolloids. The significance of these changes will depend on the scale and duration of the geological process. For example, in fractured rocks, it can be envisaged that natural geological processes, such as glaciations, would have a rapid impact on hydrological regimes (e.g. development and changes in fracture location and interconnectivity; sub-glacial penetration of water) and introduction of new nutrient sources for microbial growth (e.g. Anderson et al., 2011, Humphreys et al., 2010, Behrends et al., 2012). In other environments where groundwater advection is the main process of transport (in lower strength rocks and evaporates, transport is likely be diffusive), microbial activity will be generally low due to the small pore size, poor pore interconnectivity and resulting restricted availability of water and nutrients (Humphreys et al., 2010; Stroes-Gascoyne et al., 2007; Wersin et al., 2011). Thus it is likely that long-term natural geological processes (e.g. a succession of glaciation events) would be needed to alter rock hydrological properties impacting on indigenous microbial population activity.

Specific work on unperturbed Opalinus Clay at the Mont Terri Rock Laboratory also suggests that physical disturbances in the rock structure would provide space, water and nutrients that could revive dormant organisms. The resulting effect on porewater chemistry (e.g. pH and Eh), may affect the mobility of radionuclides, although these effects will be spatially and temporary limited because of the large buffering capacity and diffuse properties of the clay in this particular environment (Stroes-Gascoyne et al., 2007, Wersin et al., 2011). Additionally, when surface water containing oxygen encounters the “stationary” groundwater system at depth, there is an increase in microbiological activity. This results in oxygen depletion, transformation of organic material to carbon dioxide and the formation of biofilms. The time scale for complete removal of the oxygen in ‘typical’ fractures was estimated to be in the order of a few days (Pedersen, 2005). There is also evidence to show that the introduction of surface oxygenated water into deep environments has potential to re-oxidise reduced radionuclide species such as uranium rendering them more soluble and therefore more mobile (Hallbeck and Pedersen, 2008). Exogenous organisms would also be introduced into deep groundwater which could respire oxygen and potentially colonise these environments.

Microbial processes that increase gas production can also impact on the localised environment due to increases in pressure and fracturing of the rock. In terms of transport, migration rates in groundwater may also increase, for example, in gas-phase release of radionuclides where microbial metabolism incorporates ^{14}C and tritium into metabolic products such as hydrogen, hydrogen sulphide and methane (Bass et al., 2002). In addition, microbes can also affect the physical and chemical condition of the host rock, i.e. colonisation by bacteria could significantly enhance weathering of the host rock (Coombs et al, 2010; Ehrlich and Newman, 2009). As a result of increased microbial weathering, cracks and fractures within the host rock may be enlarged, which allows increased groundwater flow and gas transfer (Stroes-Gascoyne et al., 2007).

5.7 DIAGENESIS

Diagenesis is normally applied to sedimentary rock systems but the geochemical and rock-water interaction processes involved may be similar in crystalline or fractured rock systems as well. It is broadly defined as the cumulation of chemical, physical or biological changes taking place in a sediment between its deposition and the completion of lithification or cementation, and before the onset of metamorphism (Frey, 1987). Diagenetic processes are often slow and occur over long timescales, and some diagenetic processes, traditionally considered important in the evolution of aquifers, and hydrocarbon reservoirs and caprocks, may not significantly affect the geosphere or biosphere over the 1 million year time frame under consideration in the context of radioactive waste disposal.

Diagenesis can be divided into Eodiagenetic (early), Mesodiagenetic (burial) and Telodiagnetic (late) processes. There is of course a continuum between these diagenetic “stages”. Mineralogical and geochemical changes associated with these classes will be of varying significance.

Eodiagenetic or Early Diagenetic processes affect sediments in the depositional environment during deposition (syn-depositional) and before the sediments are buried under any significant overburden.

Mesodiagenetic or Burial Diagenetic processes modify sedimentary rocks during the prograding burial to deep burial stage of basin evolution. Mesodiagenesis involves interaction with warm to hot fluids, generally of increasing salinity (up to brine) with increasing depth of burial. It is often, but not always, associated with the formation of reduced mineral species such as pyrite, marcasite, ferroan carbonate cements (e.g. ferroan calcite, ankerite, ferroan dolomite, siderite) and ferroan clay minerals (e.g. chlorite, chamosite, ferroan illite).

Classic accounts of the mesodiagenetic alteration of mudrock burial sequences have been described from regions such as the US Gulf Coast, in which progressively increasing temperature and pressure produce changes in the clay mineralogy from smectite, through mixed-layer illite-smectite and chlorite smectite, to illite and chlorite clay assemblages (e.g. Hower et al., 1976; Boles and Franks, 1979). These mineralogical transformations can also be recognised in sandstones, which may also be subject to additional rock water interactions. Such thermal related effects are also observed in contact with metamorphism and hydrothermal systems, and may provide analogues for the alteration of bentonite buffer materials affected by heat-generating wastes.

Mesodiagenesis also produces changes in the sedimentary organic matter. This includes kerogen maturation and the generation of volatile hydrocarbons. The behaviour of natural organic matter undergoing diagenetic alteration may have relevance to the potential behaviour and stability of organic compounds (e.g. cellulose degradation products, $^{14}CH_4$) that may be released from a GDF.

In the context of a GDF, both eodiagenetic and mesodiagenetic processes which may have affected the host rocks will be geologically old processes, and unlikely to be relevant in terms of these processes directly influencing the future mineralogical and geochemical evolution of a GDF and its environment over the next 1Myr timescale. However, mineralogical alteration during these stages of diagenesis (and analogous processes in fractured crystalline rocks) may be very important in that it may pre-dispose the rock mass towards any subsequent late-stage telodiagenetic modifications, weathering or rock-water interaction and alteration. Eodioagenesis and mesodiagenesis can result in the precipitation of resistate mineral cements (e.g. quartz, feldspars) or more reactive, soluble mineral cements (e.g. carbonate minerals, anhydrite, gypsum, halite), which impact on porosity and permeability (e.g. Schmidt and McDonald, 1979; Burley, 1984; Burley and Kantoriwicz, 1986; Bath et al., 1987; Strong and Milodowski, 1987). This may influence both the present-day distribution of flow paths, and the way in which the flow paths and permeability may respond to future changes in groundwater chemistry. It has been long recognised that mesodiagenesis is often associated with major mineralogical, porosity and permeability changes in hydrocarbon reservoirs and aquifers.

These processes may also be important in determining the mineralogical and geochemical properties of flow-path surfaces. For example, diagenesis may control the precipitation of grain-coating clays or oxides that may subsequently influence sorption processes; and the formation of redox-sensitive phases - both oxidised species and reduced species, depending on the lithology and sedimentary environment. In turn, this may influence radionuclide interactions and migration through the rock mass.

Eodiagenetic and mesodiagenetic effects in the present-day rock mass should be predicable and the distribution of their impacts can be determined from detailed site investigation. They will potentially be influenced by lithology, and may be site-specific - controlled by the local and regional geological history.

Telodiagenesis or Late Diagenesis refers to processes that take place following uplift and/or the subsequent erosion of the rock sequence. It occurs at shallow (near-surface) to moderate depths and at low temperature. During this stage the rock sequence may be affected by meteoric water invasion during uplift, which is accompanied by the displacement or flushing of connate porewaters or mesodiagenic saline basinal fluids by dilute or fresh water. Telodiagenesis therefore forms a continuum with weathering processes.

Telodiagenesis is often associated with oxidation reactions, commonly with the formation of secondary iron and manganese oxyhydroxides and by the leaching and dissolution of soluble and reactive mineral species (e.g. Burley, 1984; Bath et al., 1987; Strong and Milodowski, 1987; Milodowski et al., 1998, 2002). Major porosity rejuvenation and enhancement commonly resulting from mineral dissolution (e.g. dissolution of evaporite cements such as halite, anhydrite, gypsum; carbonate mineral dissolution, silicate mineral dissolution), and authigenic (secondary) clay mineral formation (e.g. kaolinite) may be important. The importance of secondary porosity formation during telodiagenesis for reservoir development has long been recognised by the hydrocarbons industry (Schmidt and McDonald, 1979; Bjørlykke et al. 1992; Burley, 1984; McCaulay et al. 1994). In contrast, the significance of telodiagenetic mineral dissolution and secondary porosity formation in controlling fracture flow pathways is overlooked in many hydrogeological investigations in fractured rocks. The tendency in fractured rock environments is to assume that the structural and stress regime are the dominant control on fracture porosity and fracture flow (Sathar et al. 2012).

There are a number of processes that will influence telodiagenesis, including changes in sea-level and climate change, which may be important. In particular, changes to groundwater flow paths and hydrochemistry during glacial and permafrost phases could introduce dilute groundwaters and subsequent ‘weathering’ processes to greater depths than would be prevalent during temperate climatic phases. This may lead to alteration of, for example, fracture wall rock mineralogy that in turn may change the properties of parts of the geosphere with consequent impact on the biosphere. Such rock-water interactions will not be confined to sedimentary rocks.

Telodiagenetic processes are considered here as being of most significance to the evolution of the geosphere in the context of a GDF over a 1 Myr timescale and may be associated with tectonics, climate change and changes in hydrogeology. Major telodiagenetic alteration following Tertiary uplift is recognized in the UK: as for example, in both Permo-Triassic sedimentary cover rocks and fractured Palaeozoic basement rocks in west Cumbria (Bath et al., 2000; Milodowski et al., 1998, 2005), and in the Permo-Triassic aquifer in the East Midlands (Bath et al., 1987).

5.8 PALAEOHYDROGEOLOGY

5.8.1 Palaeohydrological Concepts In Relation To Climate Change

The main environmental changes in northern European regions during the last one million years were the development of a cover by ice sheets and/or the development of permafrost during the glacial periods, fluctuating sea-level for coastal areas with possibilities of marine inundation, variations in precipitation and evapotranspiration, and topographic changes resulting from erosion, down-cutting of river courses and sedimentation (especially associated with glacial advance and retreat).

Potential impacts of these environmental changes on groundwaters would have been caused by:

- Variations in meteoric recharge;
- Inflow of meltwater influenced by high sub-glacial pressure;
- Reduction or cessation of recharge through permafrost;
- Inflow of seawater and down-flow of this water due to density contrast; and
- Isostatic depression of the rock mass due to ice loading, and geomechanical effects on fracture permeability/porosity and matrix storage due to ice loading and other neotectonic processes.

In contrast, southern Europe did not experience either glaciation or permafrost, and variations from semi-arid to temperate/humid conditions would have caused fluctuations in water table position and salinisation of shallow groundwaters.

The stability of groundwater conditions is one of the most important safety requirements for a GDF, because the chemical composition of water and the rate of water movement are key factors influencing the reliability of containment in a GDF, the mobility of radionuclides, and the rate of their leakage back to the surface. Palaeohydrogeological investigations must address how this stability can be assessed with respect to changes in climate (or other drivers of the hydrogeological system) by investigating past fluid movement. Key issues are:

- What evidence is there that “going underground” eliminates the extreme conditions that storage on the surface would be subjected to in the long-term?
- How can the additional stability and safety of the deep geosphere be demonstrated with evidence from the natural system (i.e. natural analogue observation)?

Figure 40 summarises the generic palaeohydrogeological concepts of the interaction between changing glacial, periglacial and temperate climate states and deep groundwater systems, and their impacts on groundwater geochemistry and flow path evolution. This shows that changing climate state can potentially affect the movement and position of interfaces of bodies of water of different chemical composition. Climate change will impact on:

- **Groundwater recharge.** More arid conditions will reduce recharge and consequently result in the upward movement of the interface/mixing zone between shallow fresh groundwater and deeper saline groundwater. Conversely, increased rainfall will increase freshwater recharge and consequently drive the movement of the interface downwards.

  The development of a permafrost barrier will reduce surface recharge, and consequently the shallow fresh groundwater and deeper saline groundwater interface/mixing zone may migrate upwards.

- **Hydraulic gradients.** Hydraulic gradients may change as a result of:

  - Change in sea-level, which affects the base level of the hydraulic system;
  - Glaciations and deglaciations, which influence the hydraulic head; and
  - uplift and erosion.

- **Groundwater composition:** Freezing of groundwater associated with the development of permafrost may increase salinity and lead to the formation of residual brines. This may impact on the stability of GDF materials (e.g. corrosion behaviour) and radionuclide complexation and migration, if the brine reaches GDF depth.

  The resultant development of a body of more dense brine overlying deeper less saline groundwater may result in gravity-driven groundwater flow and salinity "overturn". Highly oxidising groundwaters derived from the melting of glacial ice may be driven to GDF depth by the increased hydraulic head induced by glacial cover. This will potentially affect the corrosion of waste canisters and increase the solubility and mobility of redox-sensitive radionuclides.

  The movement of different bodies of groundwater will result in mixing of waters of different composition within the geosphere. This may change mineral saturation and could result in mineral dissolution and enhanced porosity, or the precipitation of new minerals that may lead to sealing up of flow paths.

Understanding spatial and temporal variations in groundwater salinity is one of the key topics of palaeohydrogeology. Palaeohydrogeology also considers the possibility that specific chemical parameters such as redox and pH might have changed over time due to external factors. For example, it has been suggested that the redox condition of groundwater might have been much more oxidising during glaciations due to the penetration of oxygenated water to greater depths than at other times (Glynn et al., 1999), although that hypothesis has been countered by evidence that rock has adequate redox-buffering capacity (Guimera et al., 1999; Banwart et al., 1994; Tullborg, 1999). These variations are related to changes in driving forces for groundwater movements in the past. Another parameter that links climate and groundwater is the stable oxygen isotopic composition which relates the origin of water as meteoric recharge to the temperature at the time of infiltration.

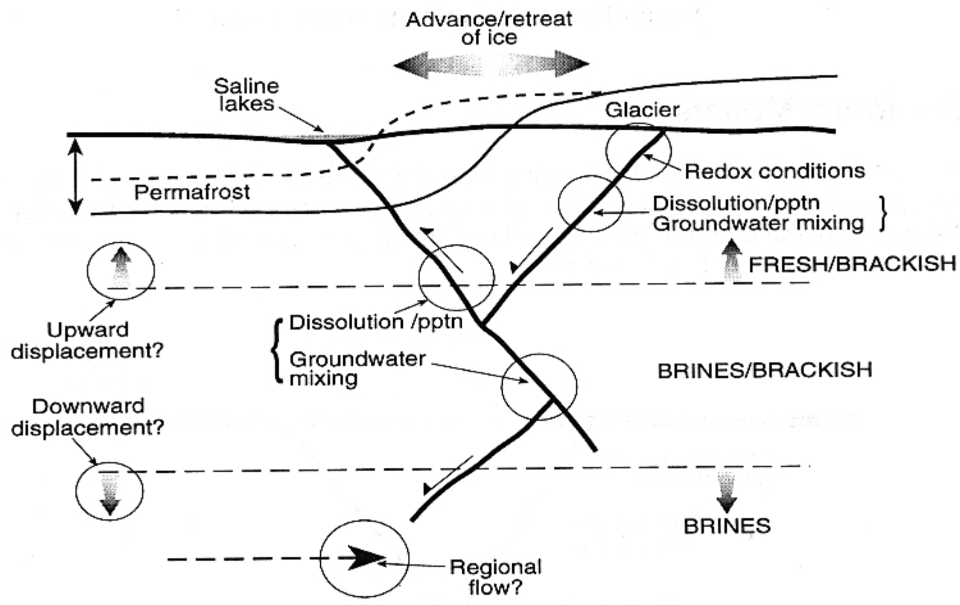


Figure 40. Illustration of the generic concepts of the potential interactions between changing glacial, periglacial and temperate climate states and deep groundwater systems.

5.8.2 Palaeohydrogeological Studies in relation to the Geological Disposal of Radioactive Wastes

Several previous studies have been undertaken in Europe and elsewhere to specifically address understanding the impact of climate change on deep groundwater systems associated with a GDF environment. These include the European Union EC 4th Framework EQUIP (Evidence from Quaternary Infills for Palaeohydrogeology), PAGEPA (Development and Testing of Models for Climate Impacts on Groundwaters) and PHYMOL (PaleoHYdrological Study of the Mol site), and European Union Framework V PADAMOT (Palaeohydrogeological Data Analysis and Model Testing) project (Blomqvist et al., 1998; Bath et al., 2000; Boulton et al., 2001; Milodowski et al., 2005 (and references therein)). These are summarised by Bath et al (2003) as part of the European Union Framework V PADAMOT Palaeohydrogeology Project. Because the Nirex investigations at Sellafield included one of the only major UK studies of palaeohydrogeology this section draws largely on the results of that study and the subsequent European Union funded projects EQUIP and PADAMOT noted above.

To build confidence in assessments of future subsurface change based on climate forecasts, the capacity to reconstruct subsurface responses to past surface climate change was tested by SKB through a palaeohydrogeological research programme (Boulton et al., 2001). As a major part of the research programme, a time-dependent, thermo-mechanically coupled model of ice sheet behaviour was developed and a large number of model simulations were carried out. The model was able to predict the internal temperature and velocity fields of the ice sheet, the temperature field and the isostatic response of the underlying bedrock, subglacial and proglacial permafrost extent and the subglacial melting rate. By connecting the glaciation model to models of groundwater flow and rock mechanical impact, the response of the subsurface to climate change was investigated. The progress made within the palaeohydrogeological programme was regarded as essential for the qualitative and quantitative descriptions of environmental effects and subsurface impact made by Boulton et al. (2001b).

The palaeohydrogeologcal study of the Sellafield site, west Cumbria has focused on the record of past events preserved in late-stage calcite mineralization lining pores and fractures in fractured Palaeozoic basement rocks and overlying Permo-Triassic sedimentary strata (Bath et al., 2000; Milodowski et al., 2005). On the basis of rigorous and detailed petrological analysis, these authors recognized a sequence of nine broad mineralization events (ME1 to ME9) and, on the basis of analysis of mineral distribution, systematic variations in crystal morphology in relation to groundwater chemistry, fluid inclusion geochemistry, stable isotope studies and strontium isotope analyses, they clearly demonstrated a very close relationship between the youngest generation (ME9) of calcite mineralization and modern deep groundwater at the Sellafield site. It was also found that the distribution of this ME9 calcite in fractures in cores could be successfully used to identify and predict the location of potential flowing features in the boreholes. Limited uranium-series dating indicated that the calcite is, at least in part, Quaternary in age.

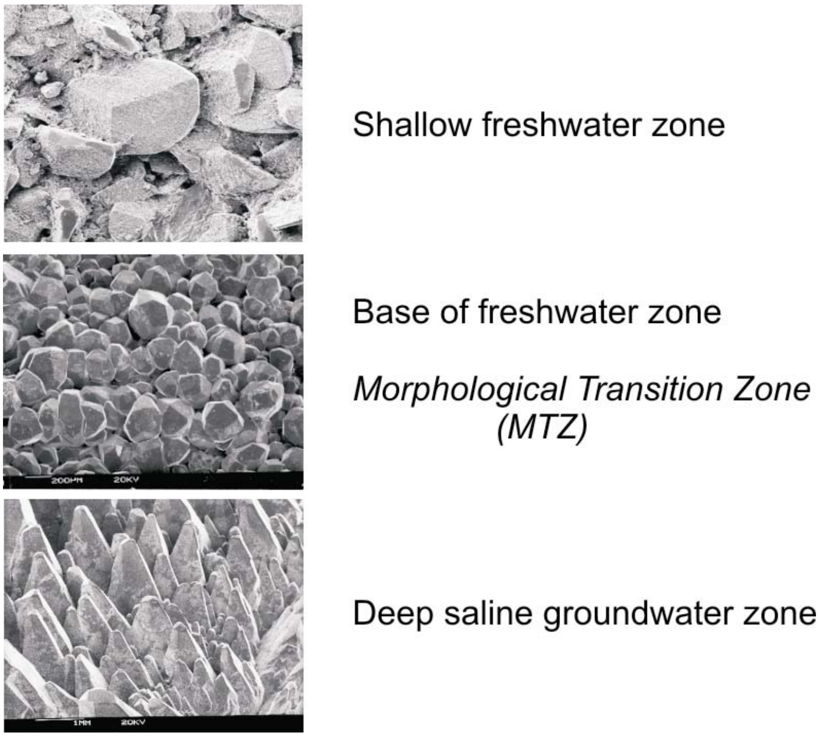


Figure 41 . SEM images illustrating the different morphological characteristics of late (ME9) calcite at Sellafield: top – short *c*-axis or ‘nailhead’ crystal form characteristic of the shallow freshwater zone; middle – equant crystal form characteristic of the deeper freshwater zone; bottom *c*-axis elongated ‘scalenohedral’ or ‘dog-tooth’ crystal form characteristic of the deeper saline groundwater zone (from Milodowski et al., 2005).

Observation of crystal morphology showed that calcite in the freshwater zone characteristically formed short c-axis crystals (‘nailhead’ calcite), whereas in the saline groundwaters the ME9 calcite was characterized by c-axis elongated crystal forms (‘dog-tooth’ or ‘scalenohedral’ calcite) (Figure 41). A systematic pattern of crystal morphology variation was observed in all of the boreholes examined, showing a progressive change from short c-axis crystals, through equant crystals to elongated c-axis crystals as present groundwater salinity increases with depth. The interval over which the crystal morphology changes, the ‘Morphological Transition Zone’ (MTZ), occurs just above the Saline Transition Zone (STZ) between the shallow freshwater zone and the deeper saline groundwater (arbitrarily defined on the basis of electrical logs at approximately 3500 to 6000 mg/L chloride). The spatial relationship between the MTZ and the STZ is maintained across the site (Figure 42). By examining the change in morphology during crystal growth, it could be demonstrated that the interface between the shallow freshwater regime and the deeper saline groundwater had progressively moved downwards with time. However, the scale of movement of this transition zone was indicated to be between 50-150 m throughout the time that the calcite has been forming (possibly initiated following Tertiary uplift). This implies that the salinity at GDF depth has probably changed little as a result of Quaternary climate change.

Detailed stable (C, O) isotope analyses of the growth-zones in the ME9 suggests that some zones in the calcite in the deep Sellafield groundwater system must have precipitated in equilibrium with groundwater containing a significant component of cold-recharged groundwater (which is different to the present-day groundwater). This was even observed to depths of 1527 m, and may indicate that glacially-recharged water may have penetrated to considerable depth in the past. However, high-resolution microchemical analysis of the calcites using electron probe microanalysis, advanced laser-ablation-inductively-coupled plasma mass spectrometry and ion microprobe techniques have shown that the deep groundwater calcites have remained ferromanaganoan throughout their growth history (i.e. redox-sensitive Fe and Mn are present as reduced species) and that redox-sensitive cerium has behaved geochemically identical to the other trivalent rare earth elements, and thus has not been influenced by oxidation to Ce4+. This shows that even if glacially recharged groundwater has penetrated to GDF depth in the past it has had little effect on the redox conditions in the deep groundwater system. The only exception to this is calcite within the shallow near-surface freshwater Permo-Triassic aquifer, where these elements have been affected by oscillations in redox behaviour. This implies that the rock mass has been well-able to buffer the redox of any oxidizing glacially-derived groundwater to reducing conditions (relative to the Fe2+/Fe3+, Mn2+/Mn3+ and Ce3+/Ce4+ redox couples). Significant oxidizing conditions (Eh exceeding +160 mV), as indicated by marked Cerium depletion in the ME9 calcite, is only observed episodically in the later stages of calcite growth (non-luminescent, non-ferroan non-manganoan calcite) above the MTZ. There is no mineralogical evidence of strongly oxidizing groundwaters ever having penetrated below the current STZ during the Quaternary.

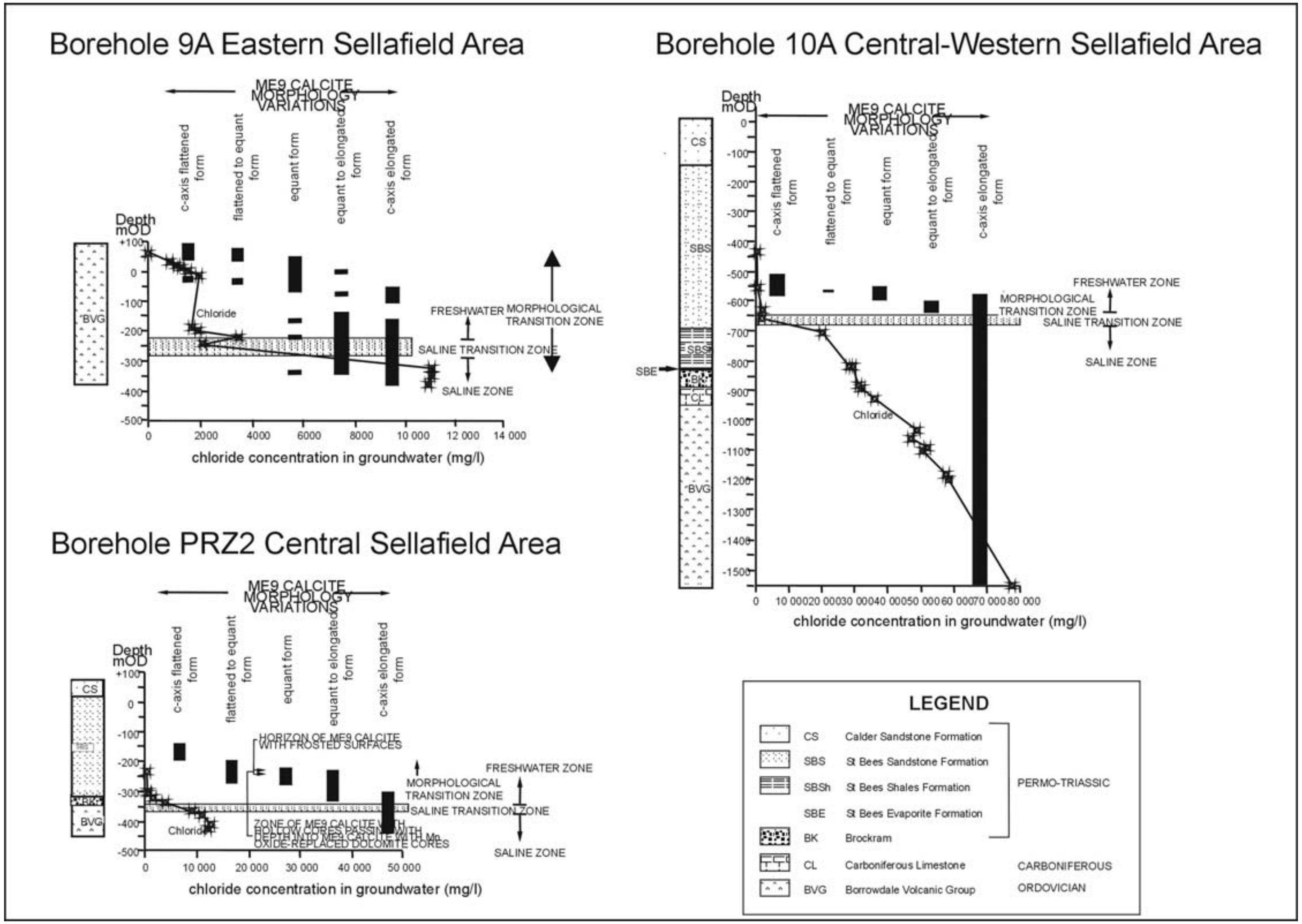


Figure 42. Distribution of morphological types of late (ME9) calcite compared to variations in present-day groundwater chemistry for Sellafield Boreholes 9A, 10A and PRZ2 (from Milodowski et al., 2005).

Petrological observations from Sellafield show that both the shallow and deep groundwater flow paths have evolved over time, and have been affected by telodiagenetic modification (Milodowski et al., 1998, 2002, 2005). Much of the fracture porosity has been demonstrated to be secondary in origin, having been re-juvenated by the dissolution of geologically old vein minerals; principally anhydrite and dolomite-ankerite (and also calcite at shallow levels) belonging to older Palaeozoic and Mesozoic mineralization (cf. also Milodowski et al., 1998, 2002, 2005). In this respect, the fracture porosity has a similar origin to a large proportion of the intergranular matrix porosity of the Permo-Triassic (St Bees Sandstone and Calder Sandstone formations) sandstone aquifer, which has been rejuvenated by the dissolution of diagenetic anhydrite, dolomite and calcite cements). Secondary porosity has long been recognised as an important factor in controlling the hydrogeological properties of sandstone aquifers and hydrocarbon reservoirs (Schmidt and McDonald, 1979). However, its role in studies of fractured groundwater systems is seldom considered (e.g. Sathar et al. 2012), and the distribution of fracture porosity is typically modelled as being related solely to tectonic structure. The PADAMOT and earlier studies show that both the palaeo, and modern, groundwater flow system progressively evolves from east to west across the site. Early anhydrite, and to a lesser extent dolomite–ankerite mineralisation, has been leached from fracture fillings, in the Borrowdale Volcanic Group (BVG) basement and Permo-Triassic cover rocks, to the east of the site as a result of invasion by dilute meteoric water (with more dissolution) in the recharge area. Similarly, anhydrite and carbonate cements are dissolving in the shallow Permo-Triassic aquifer. The enhancement of secondary porosity is progressively developing westwards and down-gradient from the recharge area as dissolution proceeds. As a result of this process, the porosity/flow pathways in the east of the site are older and more evolved, and they progressively become younger and less evolved westwards. This evolution is reflected in the growth fabric of the late-stage ME9 calcite mineralization, which precipitated from the groundwater as it became supersaturated with respect to calcite down-gradient, or as invading meteoric water mixed with background water in the host rock. Overall, this rock-water interaction process is potentially increasing the porosity in parts of the groundwater system; the void space created by the dissolution of geologically-old hydrothermal fracture mineralization (and the diagenetic cement in the Permo-Triassic strata) is generally greater than the loss of pore space due to the precipitation of new ME8 and ME9 minerals. However, localised fracture sealing by late-stage minerals is evident in the BVG, with hairline fractures becoming healed by scaly ME9 calcite. This example has demonstrated that interpretation of palaeohydrogeological information can make a valuable contribution to the development of the understanding of the evolution of a geological environment which would be needed to support a safety case for the geological disposal of radioactive waste.

This type of telodiagenetic modification to groundwater flow systems is also evident elsewhere in other areas of the UK. Bath et al., 1987b observed the same impacts of Tertiary uplift and Quaternary meteoric groundwater invasion in the East Midlands Triassic aquifer. Similar telodiagenetic alteration is seen in other Permo-Triassic basin areas in the UK (e.g., Burley, 1984; Strong and Milodowski, 1987).

5.8.3 Other Palaeohydrogeological investigations

In a study of mineral springs in mid-Wales, Edmunds et al. (1998) found that a portion of the water discharged in these springs was Pleistocene in age and originated from a depth of several hundred metres below the ground surface. This led Edmunds et al. (1998) to suggest that groundwater at this depth had been affected by glaciations, possibly the greater depth of recharge penetration during periods of glacial melting and lower sea-levels.

By studying the evolution of a coastal aquifer in north-west Portugal during the Late Pleistocene and present day using evidence from chemical and isotopic data, Condesso De Melo et al. (2001) found that radiocarbon ages indicated a smooth gradient across the aquifer, implying continuous flow during the late Pleistocene and Holocene. This suggests either that there was continuous flow during these long time periods or that changes in recharge resulting from major climate change have been 'smoothed' by other processes such as the slow rate of groundwater movement or the mixing of groundwaters of different age. However, it was found that noble gas ratios indicate that the mean annual air temperatures were lower by 5-6°C at the LGM, showing that some effects could be identified. In contrast to most areas with continental palaeowaters, the environmental isotopes indicate enrichment (0.8-1.0‰ in $^{18}0$), interpreted as reflecting the composition of the oceans at the time of the LGM at this maritime site. As most palaeowaters have a reduced ^{18}O signal, this illustrates the difficulty of interpreting ^{18}O signatures in coastal environments. This is confirmed by Loosli et al., (2001) who state that, "Indications are that the more positive ^{18}O value of ocean water during the Pleistocene dominates in the more westerly European countries over the negative $\delta^{18}0$ shift during cooler conditions".

Hinsby et al. (2001) found that Pleistocene age groundwaters, identified on the basis of stable isotopes, noble gases and corrected ^{14}C values, are present below the island of R¢m¢, off the coast of Denmark. Flow modelling results indicated that Pleistocene groundwaters were emplaced at depth within a sand aquifer known as the Ribe Formation under low base-level conditions that prevailed throughout the late Pleistocene - near the coast these waters are essentially isolated from the present flow system and Pleistocene freshwater may be present offshore. This shows how the much lower sea-levels (up to -150 m) associated with glaciations can affect the hydrogeology of an area, changing the patterns of groundwater movement. It is worth noting that most of the Irish Sea is shallower than 100 m (Figure 43)

Edmunds (2001a) used a range of specific indicators, including 3H, 3H/3He, 85Kr, chlorofluorocarbons and pollutants to recognize the extent to which waters from the modern (industrial) era have penetrated into UK aquifers, often replacing the natural palaeogroundwaters. It was found that at selected locations in the UK an age gap can be recognized indicating that no recharge took place at the time of the last glacial maximum.

The marked climatic changes that occurred during the Late Quaternary and Holocene had a significant impact on the evolution of the groundwater systems at and near the English coastline (Edmunds et al., 2001). In the Albian sands, near Worthing, freshwaters dating to 7 kyr BP are found at a depth of 450 m showing that recharge occurring 7,000 years BP has penetrated to a significant depth. In the East Midlands Sherwood Sandstone aquifer, fresh water is found to a depth of 500 m showing continuous geochemical evolution probably over a period of 100 ka, although an 'age gap' of between c. 20 and 10 kyr BP corresponds to permafrost cover. It was concluded that lowered sea-levels and the emergence of a much larger landmass over most of the past 100 kyr in places ensured deeper groundwater circulation in the vicinity of the modern coastline.

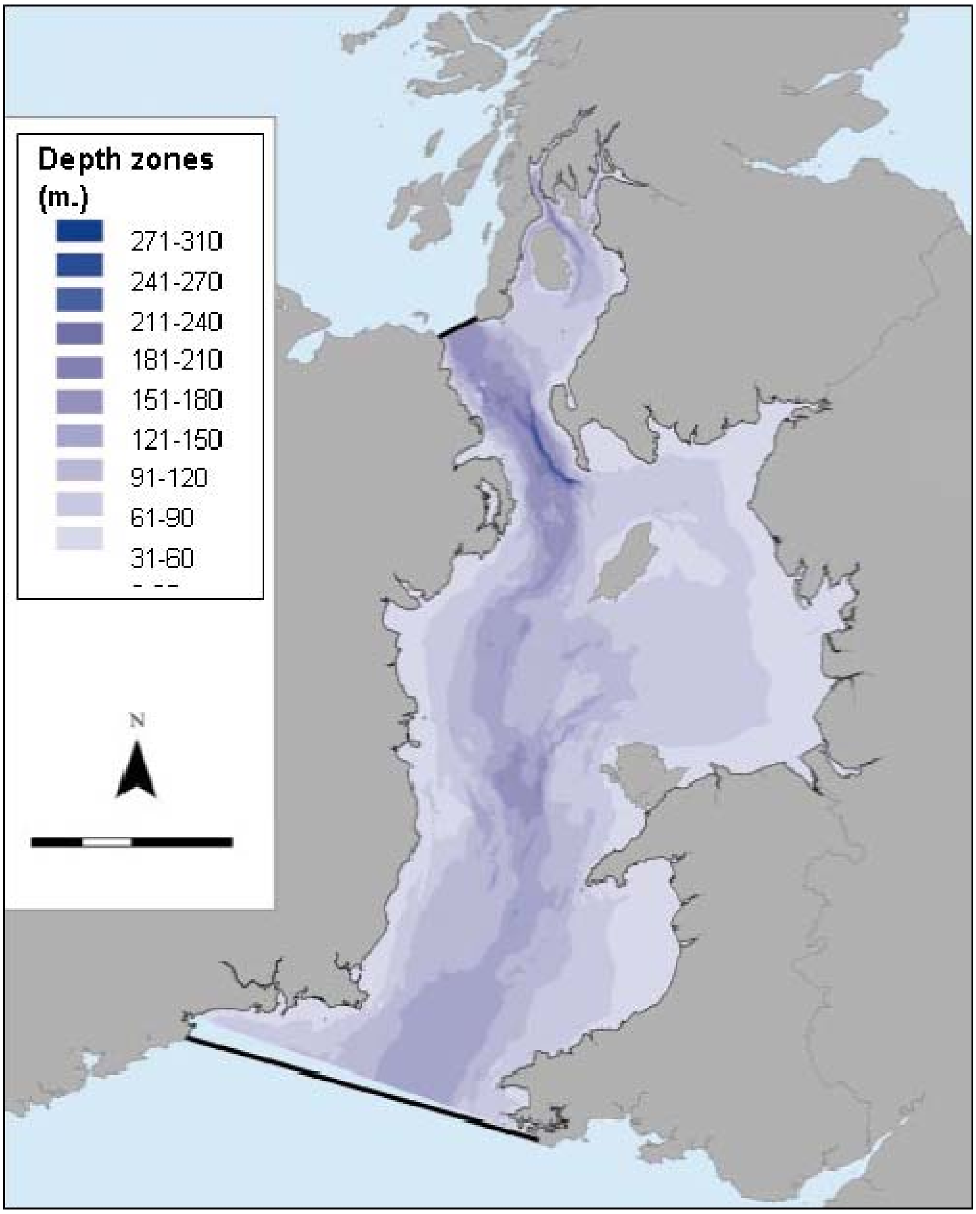


Figure 43. Irish Sea processed bathymetry (copyright BGS/NERC - http://jncc.defra.gov.uk/pdf/irish_finalreport(maps).pdf).

From a review of evidence from geophysical borehole logging techniques, Buckley et al (2001) concluded that sea-level change in response to Pleistocene glaciations and deglaciations is a major control on the salinity of groundwaters and on the development of permeable flow routes in coastal aquifers.

Edmunds et al. (2001b) state that 18 kyr BP, sea-levels around the Iberian Peninsula were as low as 150 m below those of the present day. This would promote groundwater to deeper levels than the present day. This significant lowering of the sea-level would work together with the enhanced recharge and possibly fracture opening associated deglaciation.

Vaikmae et al. (2001) state that during the last glacial maximum, recharge is likely to have ceased over much of permafrost-covered Europe, although shallow groundwater recharge from meltwater (generated by the geothermal gradients) could have taken place beneath the ice where pressure relief through tunnel valleys may have occurred. Modern recharge could have started as early as 13 ^{14}C kyr BP, but was probably interrupted by the Younger Dryas between 10 and 11 kyr BP. For most of coastal Europe, however, the most significant impact on groundwater circulation is likely to have been the lowering of sea-levels that drained large areas of the shelf, such as the North Sea and the English Channel. The Irish Sea floor is part of the shelf and would also therefore be subject to draining or see the emergence of groundwater springs and seepages. What are now marine discharges could therefore become terrestrial discharges.

Edmunds and Milne (2001) describe the PALAEAUX project which concentrated on the processes that have had a major influence on groundwater flow over the past 20 kyr years in European coastal aquifers. The most dramatic events that influenced groundwater flow during this time were global sea-level rise and climate change associated with continental-scale glaciation (Harrar et al., 2001). Permafrost and glaciation impacted several of the aquifers studied in the PALAEAUX project. It is noted that the PALEAUX study assumed that aquifer properties have remained constant over the past 20,000 years. Can glaciation affect aquifer properties? Harrar et al. (2001) showed through modelling that the interface between sea water and freshwater will move inland following sea-level rise over a period of several thousand years. It will therefore move off-shore during sea-level fall. As the position of the interface is likely to control the position of groundwater spring discharges, it is possible that these changes in the position of the interface will affect the location of groundwater discharges and possibly deep pathways and the velocity of groundwater movement at or towards GDF depths.

Numerical simulations of groundwater flow beneath continental-scale ice sheets in northern Europe indicate that during glacial periods major pressure pulses were driven through aquifer systems, resulting in high groundwater velocities and the complete reorganization of groundwater flow systems (Boulton et al. 1995; van Weert et al. 1997). These simulations also indicate that proglacial permafrost served to maintain high subsurface fluid pressures through a reduction in the bulk transmissivity of shallow aquifers and inhibition of the upward discharge of groundwater in regions in front of the glaciers. However, permafrost may have been discontinuous in discharge zones due to the thermal effects of upwelling groundwater (van Weert et al. 1997). Estimates indicate that, during the last glaciation, recharge to flow systems by meltwater production at the base of ice sheets was less than, or about the same as, present recharge rates to the regional, deep-seated aquifers (Harrar et al., 2001). Harrar et al. (2001) modelled the effects of glaciation on a large regional-scale aquifer in Belgium. This suggested that (i) high pressures under the ice sheet penetrated to depth throughout the entire aquifer system, (ii) in the deep part of the aquifer system, the propagation of the high-pressure front caused an upwelling of the deep groundwater towards the south (discharge zone) and the upper model layers and (iii) significant changes in salinity require a thick ice sheet (at least a few hundreds of metres) over a sufficiently long time (a least several ten thousands of years).

Harrar et al. (2001) make the following conclusions in relation to the response of open-type aquifers (i.e. discharging to the sea) to sea-level rise:

- The pressure effects caused by changes in sea-level propagate through the aquifer system at rates dominated by aquifer hydraulic conductivity and storage. Hydraulic heads stabilize relatively quickly: characteristic times for pressure changes to propagate through the system are in the order of 10 to 100 years. Present hydraulic heads are probably close to a steady state;
- Flow responds less quickly than pressure, and is controlled by aquifer thickness and porosity and aquitard permeability and thickness;

- Offshore regional head gradients and fluxes become very small and diffusion may be an important process governing interface movement;
- Encroachment of the saltwater interface associated with sea-level rise takes a long time (of the order of 10,000 years); and,
- The position of the saltwater interface is not at a steady state.

With respect to the effects of glaciation and permafrost, Harrar et al. (2001) made the following conclusions:

- Glaciation alters the flow patterns within an aquifer system;
- The effect of the glacier is not always to push freshwater deeper into the formation: the geometry of the aquifer in relation to the glacier is of great importance;
- The effect of the ice sheet is different if it overlies the aquifer outcrop area directly;
- Permafrost reduces direct recharge to the aquifer system resulting in a decrease in hydraulic heads and aquifer fluxes;
- Permafrost may alter groundwater flow paths by reducing natural surface discharge;
- If a significant depth of formation is frozen due to permafrost then the hydraulic properties of the aquifer system are changed;
- As little is known about the precise position and extent of glaciers and permafrost it is difficult to make reliable predictions on the effect they have on recharge rates and flow directions in the aquifers.

Harrar et al. (2001) also concluded that it is not possible to make any useful general statements about the effects of glaciation and permafrost on aquifers as the relative geometry of the aquifer and glacier are very important.

5.9 WEATHERING AND EROSION: POTENTIAL IMPACTS ON A GDF OVER THE NEXT ONE MILLION YEARS

The major issues related to erosion and weathering are relatively shallow and in the majority of cases will have little effect on a GDF placed within a depth of between 200 and 1000 m over the next one million years. Associated infrastructure, such as backfilled access shafts and drifts, will be affected in the near surface environment by weathering and erosion processes but are not considered further here.

The denudation rates of most hard rock types under non-orogenic conditions is less than 100 m Myr^{-1} and in most reported cases <50 m (Table 6 above). These denudation rates would potentially remove some of the near surface parts of the backfilled shafts and access ramps but would not impact on a GDF itself. Because of its mid-tectonic plate location the greater rates of denudation that are found in areas of high glacio-eustatic change with soft rocks, such as Kobo Peninsula in Japan, will not occur in the UK. Rocks that dissolve easily, such as limestones, halite and gypsum, have potentially much greater erosion rates if exposed to circulating groundwaters.

River incision resulting from the isostatic change in the UK after glacial re-adjustment has been found to reach around 160 m in the Thames river system over the duration of the Quaternary. Again, this may potentially remove some of the near surface parts of the backfilled shafts and access ramps but would not impact on a GDF itself though if this amount of incision occurs above a GDF at 200 m the remaining cover is significantly reduced.

It is hard to predict the dynamics of ice sheets and glaciers, the effects of which have been addressed in Chapter 4 above. The extent of such processes would depend on a number of factors including the duration and intensity of any future events and therefore secondary geomorphological processes such as river capture and the subsequent re-routing of river systems and how these will impact a GDF are equally difficult to predict.

The greatest eroding forces are those connected with the presence of ice sheets and glaciers. In upland areas, glacial overdeepening in the UK can reach depths of about 200 m, as shown by examination of U shaped valleys, hidden valleys and lochs, but has been shown to be greater (up to a maximum of about 1300 m) in Norway where ice was significantly thicker. It is likely that future glaciers would continue the modification of existing valley systems.

Water-rock interactions, including chemical weathering, have the potential to alter groundwater pathways by enhancing dissolution or deposition of minerals. This could allow changes in microbial communities and redox conditions that may affect the transport of radionuclides. However, evidence, including from the Sellafield site investigation, from fracture fills and groundwater chemistry shows that from relatively shallow depths, and certainly within 200 m, the rock mass effectively buffers deeper groundwater so that it remains reducing.

Near surface weathering causes changes to the permeability of the rock mass and to an extent is caused by interactions with groundwater. As a result of mineral alteration; dissolution; or deposition, flow paths may be enhanced or restricted, changing how groundwater flows though the rock mass. These processes may also result in changes to the chemistry of groundwater as minerals are altered, dissolved or deposited. Overall changes to the groundwater flow pathways as a result of weathering processes are likely to occur within the shallow part of a groundwater system and will have little effect on a GDF at depth. On the other hand, diagenetic and related processes occur at all stages during the geological evolution of the rock mass and will evolve with time as the geological environment changes with time. The effects are similar to those induced by weathering and included mineral alteration, dissolution or deposition and also result in changes to the chemistry of groundwater as minerals are altered, dissolved or deposited.

6 Conclusions

This review has focused on a number of geological disciplines where there may be significant effects on a GDF at depths of between 200 m and 1 km beneath the UK. In particular it has examined the somewhat inter-related disciplines relating to climate change, including glaciation, weathering and erosion, including diagenetic effects, seismics and tectonics, including uplift or subsidence. Volcanism in the UK is considered highly improbable over the next one million years (or very much longer). For the majority of the potential natural changes covered in this review it is considered that the possible effects on a GDF are likely to be minimal in all of the geological environments considered ('basement' to surface (both in lower and higher altitude terrains); 'basement' under sedimentary cover (both in lower and higher altitude terrains); mudrock (in lower altitude terrains only) and bedded evaporites (in lower altitude terrains only)) over the next one million years. These processes are not considered further in this chapter, which only covers the potentially more significant effects, even though these may be unlikely to occur. Table 12 summarises the events that may be relevant to a GDF in the UK. The majority of these will affect all of the generic geological environments consided by the RWMD and therefore these environments are not specifically noted.

Event	**Impact on a GDF if facility directly affected. Typical depth of GDF about 600 m assumed**	**Likelihood of occurrence somewhere in the UK in next 1Myr**	**Comments**
Glacial Erosion (upland)	Minor to none	Highly probable	Only a GDF close to 200m depth could be affected
Glacial Erosion (lowland)	None	Highly probable	
Permafrost	Moderate to minimal	Highly probable	Depends on depth of a GDF and depth of ground freezing
Sub-glacial fluvial erosion (upland)	Minor to none	Highly probable	Only a GDF close to 200m depth could be affected
Sub-glacial fluvial erosion (lowland)	None	Highly probable	
Fluvio-glacial outwash stream incision	None	Highly probable	
Slope failure	None	Probable	Upland areas and steeper slopes only
Sea-level rise	Minimal	Highly probable	Lower lying areas only, likely to change groundwater flow paths as base levels change
Sea level fall	Minor	Highly probable	Shallow coastal areas only, likely to change groundwater flow paths as base levels change
Frost related denudation	None	Highly probable	
Weathering	Minor to none	Definite	
Erosion (land mass lowering)	Minor to none	Definite	
Diagenesis	Minor to none		
Earthquake induced rupture (faulting)	Moderate	Probable	Highly unlikely to occur at a GDF location
Earthquake induced vibration	Minimal	Probable	Highly unlikely to occur at a GDF location
Earthquake induced deformation	Minimal	Probable	Highly unlikely to occur at a GDF location
Earthquake induced liquifaction	None	Probable	Highly unlikely to occur at a GDF location
Tsunami	None	Probable	Lower lying areas only
Tectonic related uplift/subsidence	Minimal	Probable	
Isostatic adjustment	Minimal	Highly probable	
Mineralisation	Minor	Highly probable	
Volcanism	Major	Highly unlikely	Volcanism is not anticipated in the UK over next 10's Ma

Table 12: Future natural change events and their potential impacts on a GDF in the UK.

6.1 IMPACTS RELATED TO UPLIFT AND SUBSIDENCE

Most of the factors related to uplift and subsidence will not significantly affect a GDF in any rock types in any location in the UK over the next one million years and some have been addressed above. Fracture formation is typically accompanied by mineralisation by mineral-rich fluids flowing through the rock matrix (where it can) and along the fracture planes themselves. In fact, in some geological settings fracturing does not occur without some form of mineralisation. The extent and type of mineralisation is affected by the hydrogeology and size, distribution and extent of fractures. Over the one million year time frame such diagenetic processes, which may include dissolution as well as deposition of minerals in fractures and pore spaces may enhance and/or reduce groundwater flows.

6.2 IMPACTS RELATED TO SESMICITY AND TECTONICS

Earthquake activity presents a number of hazards for radioactive waste disposal which, while the likelihood of occurring is low, could have a significant effect on a GDF if such events occurred. These include:

- Fault displacement (rupture) hazard;
- Vibratory hazard (ground shaking); and
- Secondary hazards (e.g. changes to groundwater).

Fault displacement hazard refers to the danger of physical movement along a fault plane disrupting the waste emplacement; one can distinguish between principal fault hazard (movement along the fault plane of the earthquake) and distributed fault hazard (secondary movement at some distance from the principal fault plane). Vibratory hazard concerns the possibility of damage due to strong shaking. Secondary hazard due to seismic disruption of ground water patterns is also possible.

The rupture hazard for a shallow GDF in a low seismicity intraplate region such as the British Isles is rather low because larger earthquakes tend to nucleate at mid-crustal depths. Rupture dimensions for the largest recorded earthquakes in the UK are typically of a few kilometres, so although a rupture that nucleates at depth is more likely to propagate into a region of lower strength, the potential for it to reach the surface is limited. Nevertheless, there are a number of examples of large earthquakes in intraplate areas that nucleate in the crystalline basement and rupture through to the surface, however, the magnitudes of these earthquakes are far in excess of the expected maximum magnitudes for earthquakes in and around the British Isles. No British earthquake recorded either historically or instrumentally has produced a surface rupture. Because earthquakes are likely to occur on pre-existing faults, any underground GDF should be constructed in a location where the number of existing faults is low. However, because this may not be possible, the structure should be able to tolerate expected fault displacements. Fault displacements for larger British earthquakes are likely to be in the range of several centimetres.

A number of studies have shown that in general ground motions (vibrations) at depth are less than those at the surface. In addition, several studies have documented earthquake damage to underground structures such as tunnels which generally conclude that underground structures suffer appreciably less damage than surface structures. This suggests that the shaking hazard for a buried GDF is rather less than that at the surface.

Earthquake related hydrological changes, such as creation of new springs, changes to water table levels and increased groundwater discharge have all been widely observed after a number of moderate to large earthquakes. These hydrologic changes are a response to strain caused by earthquakes, changing fluid pressures and altering hydrogeological properties such as permeability, which controls the rate of fluid flow. The amplitude of the changes may be large and can often be observed at large distances from the earthquake. These may affect the hydrogeology in and around a GDF.

Permanent changes to porosity and permeability resulting from fault rupture and deformation during an earthquake should be considered as potential hazards to a GDF.

The possibility of renewed glaciation within the lifetime of a GDF means that estimates of the distribution and rates of regional seismicity cannot be considered the same as present. Geological investigations in a number of regions have found evidence for significant post-glacial movement of large neotectonic fault systems, which were likely to have produced large earthquakes around the endglacial period. For example the 150 km long, 13 m high fault scarp of the Pårve Fault in Sweden is believed to have been caused by a series of post-glacial earthquakes. Similar evidence for post-glacial fault displacements in Scotland is known however these are likely to be restricted to metre-scale vertical movements along pre-existing faults.

A number of studies suggest that earthquake activity beneath an ice sheet is likely to be suppressed but is followed by much higher levels of activity after the ice has retreated again. Consequently, estimates of seismicity based on current rates may be quite misleading as to the possible levels of activity that could occur in the more distant future. It should be noted that the largest stress changes occur at the former ice margins, making these the most likely source region for enhanced earthquake activity. The implication for a GDF in such a region is that seismicity rates following any future glacial period may be significantly higher than at present. Given our current maximum magnitude in the UK of around 6 M_W it is not unreasonable to expect an increase in the maximum possible magnitude to 7 M_W following such an event.

6.3 IMPACTS RELATED TO GLACIATION

The depths of erosion, particularly in uplands and adjacent areas over the one million year time frame are potentially of the order of 200 metres below existing ground surface profiles. The areas where such deep erosion occurs will be localised and will be dependent on where active ice streams, major glacial meltwater drainage routes and major fluvio-glacial out-flows occur. In these areas this is likely to be mainly controlled by existing topography. The eventual depth that the valleys and channels attain will depend on factors such as ice thickness and sea level, but all are probably in the 200 m rather than 1 km depth range. The hetrogeneous pattern of glacial erosion/mass transport means that values of 'averaged or mean erosion rates for a terrain over a glacial-interglacial cycle' are relatively meaningless. A GDF at a depth of 200 to 300 m sited in an area where valley glaciers or sub- and pro-glacial streams may be focused by current topography may be exposed, or nearly exhumed, by these. At greater depths the amount of cover rock will be reduced though this may only have a modest effect on a GDF.

The combination of topography (precipitation shadow) or elevation (mountain tops are covered by less ice in an ice sheet glaciation than pre-existing valleys) will also influence ice thickness and glacier type. In upland glacial systems frost shattering of areas not covered in ice may be as effective at lowering the mountain tops as the glaciers are in deepening their beds. This process may significantly reduce the thickness of the cover over a GDF situated in a upland area depending on location and factors such as the duration of cold phases, ice thickness and ice temperature.

The depth of permafrost if prolonged periglacial activity occurs could extend beyond 200m depth (it has in the past) and perhaps as far as 1km. If such deep permafrost is achieved this may affect the engineering properties of 'soft' rocks and could lead to the development of new fracture pathways, perhaps to the surface, in more brittle formations. It also affects groundwater recharge and discharge amounts and positions. The formation of ice changes the chemistry of the remaining liquid phase concentrating dissolved salts. Permafrost could also affect the engineered elements of a GDF in similar ways, in particular the properties of clay and cement based backfill/buffer materials may be permanently or temporarily changed by permafrost conditions. Because halite is a 'dry' rock it will not be affected by permafrost, even if temperatures are reduced to below 0ºC.

Changes of relative sea level during glacial and interglacial stages could mean that a GDF site is further or nearer to the coast or even beneath the sea bed. While this may lead to coastal erosion or deposition around the surface part of a GDF (only affecting a few 10s of metres depth) it will change the groundwater flow paths by changing the base level of discharges.

6.4 IMPACTS RELATED TO WEATHERING AND EROSION

The major issues related to erosion and weathering are relatively shallow and in the majority of cases will have little effect on a GDF placed within a depth of between 200 and 1000 m over the next one million years. Associated infrastructure, such as backfilled access shafts and drifts, will be affected in the near surface environment by weathering and erosion processes but are not considered further here.

River incision resulting from the isostatic change in the UK after glacial re-adjustment has been found to reach around 160m in the Thames river system over the Quaternary, and therefore a reasonable estimate of maximum future river incision over one million years is about 65 m. This may potentially remove some of the near surface parts of the backfilled shafts and access ramps but would not impact on a GDF itself though if this amount of incision occurs above a GDF at 200m the remaining cover will be significantly reduced.

Water-rock interactions, including chemical weathering, have the potential to alter groundwater pathways by enhancing dissolution or deposition of minerals. This could allow changes in microbial communities and redox conditions that may affect the transport of radionuclides. However, evidence, including that from the Sellafield site investigations which are the most thoroughly characterised sites in the UK and from elsewhere in Europe (Milodowski et al; 2005), from fracture fills and groundwater chemistry shows that from relatively shallow depths, and certainly within 200m, the rock mass effectively buffers groundwater so that it is reducing.

6.5 IMPACTS ON GROUNDWATER

Groundwater movement is the principal mechanism by which radionuclides will eventually migrate from a GDF in all rock types. Many of the processes described in the preceding chapters are likely to have some impact on groundwater, leading to changes in flow paths, rates and groundwater chemistry that in turn may enhance or retard radionuclide migration.

Any processes that cause uplift of the land relative to sea level may lower the base level to which groundwater flows are naturally directed. This may then lead to changes in the groundwater flow paths including:

- Higher heads;
- Higher flow rates;
- Longer flow paths; and,
- Deeper flow paths.

In the opposite case, where the level of the land relative to sea level is reduced, the reverse situation is likely.

Development of ice sheets may also lead to the development of higher heads resulting in higher flow rates and possibly deeper flow paths, while permafrost may cause the diversion of groundwater flows around frozen ground. Both ice and permafrost may block pre-cold phase outflow locations, leading to diversion of ground water flows to other outlets or the development of new outlets.

Where groundwater flow paths are diverted, or develop at great depths because of any of these changes, there is potential for a GDF to be affected by the new groundwater flow regime.

Near surface weathering causes changes to the permeability of the rock mass. As a result of mineral alteration, dissolution or deposition flow paths may be enhanced or restricted, changing how groundwater flows though the rock mass.

The chemistry of groundwater will be altered by weathering, freezing and diagenetic processes and may result in the alteration, dissolution or deposition of minerals, particularly in fractures and pore spaces, and to changes in solute concentrations and changes to the redox state. Enhanced recharge by glacial melt water may lead to changes in redox conditions because of the introduction of oxygen-rich water. This is likely to be a near surface, perhaps 100 to 200 m, effect because of the buffering properties of the rock mass as a whole.

The changes in groundwater flow paths and chemistry noted above are likely to be long term or permanent. Most will happen gradually as the landscape and underlying rock mass evolves, but those induced by glacial process are likely to be geologically rapid and occur over relatively short timescales.

Given the scale of seismicity in the UK, seismic events are likely to have only transient impacts on groundwater flow. Permanent deformation may lead to increase or decrease in bulk rock permeability, mainly in the vicinity of the active fault.

The alteration of groundwater flow regimes, which may result in freshwater reaching greater depths than currently, possibly as deep as a GDF, with associated changes to the chemistry of the groundwater, has the potential to alter the geochemical, biological and mineralogical processes occurring in and around a GDF. This might result in enhanced or retarded radionuclide migration as a GDF and the surrounding environment evolve.

6.6 EVAPORITES – A SPECIAL CASE

Of the rock types that may be considered for the disposal of radioactive waste evaporites, in particular salt, are a special case because all evaporites are soluble and in unsaturated groundwater all will dissolve quickly. If any of the processes noted in the summaries above change groundwater flow patterns to the extent that evaporites are in direct contact with fresh groundwater then dissolution will occur which is likely to have a significant effect on a GDF hosted in such rocks.

6.7 SUMMARY

For the majority of processes covered in this review, the possible effects on a GDF are likely to be minimal in all of the geological environments considered over the next one million years.

There are a number of processes that may have an effect on a UK GDF, depending on location, but the likelihood of significant consequences is low. In particular the following processes have been identified as having the potential to affect a GDF if circumstances are unfavourable:

- In certain circumstances, glacial erosion could significantly reduce the cover on a GDF and thus has the potential to directly affect a shallowly sited GDF (around 200 m depth) in an area where existing glacial valleys focus future glacial flow or where sub- and pro-glacial streams cause deep downward erosion;
- Permafrost conditions may extend to several 100 m depth and could alter fracture pathways in brittle formations, affect groundwater recharge and discharge and influence clay and cement properties;
- Erosion and weathering are relatively shallow processes and, consequently, will largely affect near-surface infrastructure, such as backfilled access shafts and drifts. Weathering also has the potential to alter groundwater pathways by enhancing dissolution or deposition of minerals, which could enhance or retard the transport of radionuclides;
- The potential for seismic damage, caused by vibration and fault displacement, is low but secondary effects, acting through changes to groundwater and elevated post-glacial increases in seismicity rates, could affect a GDF. Similarly, the main risk from uplift or subsidence is through changes to groundwater flow affecting water-rock interactions and diagenesis;
- If any of the above processes change groundwater flow patterns to the extent that evaporites are in direct contact with fresh groundwater, then their dissolution will occur, which is likely to have a significant effect on a GDF hosted in such rocks.

Until specific sites have been identified as potential locations to host a GDF it is not appropriate to undertake a detailed assessment of many future changes noted above that may affect a GDF in a generic, non-site specific way. There are a number of issues relating to the formation and thawing of permafrost during cold climatic conditions that would benefit from generic research. Once potential sites have been identified and rock types are known specific factors relevant to the site can then be assessed.

7 References

Abercrombie, R. E., 1997. Near-surface attenuation and site effects from comparison of surface and deepborehole recordings. Bull. Seismol. Soc. Am. 87 (3), 731-744.

Adams, J., 1989a. Postglacial faulting in eastern Canada: nature, origin and seismic-hazard implication. Tectonophys., 163, 323-331.

Adams, J., 1989b. Crustal stresses in eastern Canada. In: GREGERSEN, S., BASHAM, P.W. (Eds.), Earthquakes at North Atlantic Passive Margins: Neotectonics and Postglacial Rebound. Kluwer, Dordrecht, 289-297.

Adams, J., 1996. Paleoseismology in Canada: a dozen years of progress. J. Geophys. Res., 101, 6193-6207.

Afifi A A, Bricker O P, 1983. Weathering reactions, water chemistry and denudation rates in drainage basins of different bedrock types: 1 – sandstone and shale. In: Webb B W (ed). Dissolved loads of rivers and surface water quantity/quality relationships: proceedings of a symposium held during the XVIII General Assembly of the International Union of Geodesy and Geophysics at Hamburg, 15–27 August 1983. Paris: International Association of Hydrological Sciences. (IAHS-AISH publication 141), pp 193–203.

Ahonen, L., Kukkonen, I., Toppi, T., Nyyssonen, M., Bomberg, M., Nousiainen, A. and Itavaara, M. 2010. Deep Life and Gases in the Outokumpu Deep Borehole: Baseline Information for Nuclear Waste Disposal in Crystalline Rock. MRS Proceedings. 1265: 197-202.

Albrecht, A., Herzog, G.F., Klein, J., Dezfouly-Arjomandy, B. and Goff, F. 1993. Quartenary erosion and cosmic-ray-exposure history derived from 10Be and 26Al produced in-situ. An example from Pajarito plateau, Valles Caldera region. Geology, 21, 551-554.

Allison, I., Bindoff, N. L., Bindschadler, R. A., Cox, P. M., de Noblet, N., England, M. H., Francis, J. E., Gruber, N., Haywood, A. M., Karoly, D. J., Kaser, G., Le Quere, C., Lenton, T. M., Mann, M. E., McNeil, B. I., Pitman, A. J., Rahmstorf, S., Rignot, E., Schellnhuber, H. J., Schneider, S. H., Sherwood, S. C., Somerville, R. C. J., Steffen, K., Steig, E. J., Visbeck, Martin and Weaver, A. J. (2009) The Copenhagen Diagnosis: Updating the World on the Latest Climate Science Elsevier, Oxford, UK, 114 pp.

Alonso, V. and Corte, A.E., 1992. Postglacial fracturing in the Cantabrian Cordillera (NW Spain). Zeitschrift fűr Geomorphologie, 36, 479-490.

Alvarez-Solas, J. and Ramstein, G. 2011. On the triggering mechanism of Heinrich events. Proceedings of the National Academy of Sciences, 108, E1359-1360.

Ambraseys, N.N. and Jackson, J.A., 1985. Long-term Seismicity of Britain, in Earthquake Engineering in Britain, Thomas Telford, London, pp. 49-66.

Amelung, F. and Bell, J.W. 2003. Interferometric synthetic aperture radar observations of the 1994 Double Spring Flat, Nevada, earthquake (M5.9): Main shock accompanied by triggered slip on a conjugate fault. Journal of Geophysical Research, 108, 2433.

Anderson, C., Johnsson, A., Moll, H. and Pedersen, K. 2011. Radionuclide Geomicrobiology of the Deep Biosphere. Geomicrobiology Journal. 28: 540-561.

Anderson, N.J. and Leng, M. J. 2004. Increased aridity during the early Holocene in West Greenland inferred from stable isotopes in laminated-lake sediments, Quaternary Science Reviews, 23, 841-849.

Andra 2005 Dossier Argile: Synthesis - Evaluation of the feasibility of a geological repository in an argillaceous formation, Reference 266VA ISBN 2-916162-00-3, 241p

Andrews, J. T., Gilbertson, D. D. and Hawkins, A. B. 1984. The Pleistocene succession of the Severn Estuary: a revised model based upon amino acid racemization studies J. Geol. Soc. London. 141, 967-974.

Andrieu-Ponel, V., Texier, D., Brulhet, J., De Beaulieu, J-L., Begeot, C. Cheddadi, R. Gandouin, E., Goodess, C., Guiter, F., Hamelin, B., Loutre, M. F., Magny, M., Masson-Delmotte, V. Paillard, D., Petit-Marie, N., Ponel, P., Pons-Branchu, E., Reille, M., Ruffaldi, P., Thorne, M. and Van-Vliet-Lanoe, B. 2007. Climate and environmental evolution scenarios for the Meuse/Haute-Marne region, France Mémoires de la Société Géologique de France. 17, 115-134.

Arrowsmith, S.J., Kendall, M., White, N., VanDecar, J. and Booth, D.C., 2005. Seismic imaging of a hot upwelling beneath the British Isles. Geology 33 (5), 345–348.

Arvidsson, R. 1996. Fennoscandian earthquakes: Whole crustal rupturing related to postglacial rebound. Science, 274, 744–746.

Assumpcao, M. 1981. The NW Scotland earthquake swarm of 1974. Geophysical Journal of the Royal Astronomical Society, 67, 577-586.

Atkinson, T.C. Briffa, K.R. and Coope G.R. 1987. Seasonal temperatures in Britain during the past 22,000 years, reconstructed using beetle remains. Nature, 325, 587–592.

Axtmann, E.V., and Stallard, R.F. 1995. Chemical weatherimng in the South Cascade Glaciar Basin: Comparison of subglacial and extra-glacial weathering, In Tonnessen, K. A., Williams, M. W., & Tranter, M. eds., Biogeochemistry of seasonally snow-covered catchments: Boulder, Colorado, International Association of hydrological Sciences, v.228, p.431-439.

Ballantyne, C. K. 1987. The present-day periglaciation of upland Britain. 113-126, in Periglacial processes and land-forms in Britain and Ireland. Boardman, J. (editor). (Cambridge University Press).

Ballantyne, C. K. 2010. Extent and deglacial chronology of the last British-Irish Ice Sheet: implications of exposure dating using cosmogenic isotopes. J. Quaternary Sci., 25, 515–534.

Ballantyne, C K, and Harris, C. 1994. The periglaciation of Great Britain. (Cambridge: Cambridge University Press.)

Ballantyne, C.K., McCaroll, D., Nesje, A., Dahl, S.O. and Stone, J.O., 1998. The last ice sheet in North-west Scotland: reconstruction and implications. Quat. Sci. Rev. 17, 1149–1184.

Bally, A.W., Bernoulli, D., Davis, G.A. and Montadert, L. 1981. Listric normal faults. Oceanologica Acta, Proceedings 26th International Geological Congress, Paris, 87-101.

Balson, P.S. and Humphries, B. 1986. The nature and origin of fissures in the East Anglian Coralline and Red Crags. Journal Quaternary Science, 1, 13-19.

Banwart, S., Gustafsson, E., Laaksoharju, M., Nilsson, A-C., Tullborg, E-L. and Wallin, B. 1994. Large-scale intrusion of shallow water into a vertical fracture zone in crystalline bedrock: initial hydrochemical perturbation during tunnel construction at the Äspö Hard Rock Laboratory, south-eastern Sweden. Water Resources Research, 30: 1747-1763.

Baptie, B., 2010. Seismogenesis and state of stress in the UK. Tectonophysics, 482 (1-4), 150-159.

Barber, K. E., Chambers, F. M. and Maddy, D. 2003. Holocene palaeoclimates from peat stratigraphy: macrofossil proxy climate records from three oceanic raised bogs in England and Ireland. Quaternary Science Reviews, 22, 521-539.

Bass, C. J., Holtom, G. J., Jackson, C. P. and Lappin-Scott, H. 2002. The Potential Impact of Micro-organisms in the Geosphere on Radionuclide Migration. AEA Technology Report. AEAT/ERRA-0329.

Bates, M. R., Briant, R. M., Rhodes, E. J., Schwenninger, J-L. and Whittaker, J. E. 2010. A new chronological framework for Middle and Upper Pleistocene landscape evolution in the Sussex/Hampshire Coastal Corridor, UK. Proceedings of the Geologists' Association, 121, 369-392.

Bath, A.H., Milodowski, A.E. and Spiro, B. 1987a. Diagenesis of carbonate cements in Permo-Triassic sandstones in the Wessex and East Yorkshire-Lincolnshire Basins, UK: a stable isotope study. In: Marshall, J.D. (editor). Diagenesis of Sedimentary Sequences, Geological Society Special Publication No. 36, 173-190.

Bath, A.H., Milodowski, A.E. and Strong, G.E. 1987b. Fluid flow and diagenesis in the East Midlands Triassic sandstone aquifer. In: Goff, J.C. and Williams, B.P.J. (editors). Fluid Flow in Sedimentary Basins and Aquifers, Geological Society Special Publication No. 34, 127-140.

Bath, A.H., Boulton, G., Mariovoet, J. and Blomqvist, R. 2000. What approach and tools do we have for understanding the past evolution of groundwater systems as a guide to future evolution for repository performance assessment? In: Euradwaste 1999 "Radioactive Waste Management Straegies and Issues" (C Davies, Editor). Proceedings of 5th European Commission Conference on Radioactive Waste Management, Disposal and Decommissioning, Luxembourg, November 1999." pp. 257-273. European Commission, Nuclear Science and Technology Series, Luxembourg. EUR 19143 EN.

Bath, A.H., Milodowski, A., Ruotsalainen, P., Tullborg, E-L., Cortés Ruiz, A. and Aranyossy, J-F. 2000. Evidence from mineralogy and geochemistry for the evolution of groundwater systems during the Quaternary for use in radioactive waste repository safety assessment (EQUIP project). Report, European Commission, Nuclear Science and Technology, EUR 19613, D-G for Research, European Commission, Brussels.

Bayer, U., Scheck, M. and Rabbel, W. 1999. An integrated stud y of the NE German Basin. Tectonophysics, 314, 285–307.

Beaumont, C., Jamieson, R.A., Nguyen, M.H. and Medvedev, S. 2004. Crustal channel flows: 1. Numerical models with applications to the tectonics of the Himalayan–Tibetan orogen. Journal of Geophysical Research, 109, B06406.

Becker, A., Ferry, M., Monecke, K., Schnellmann, M. and Giardini, D. 2005. Multiarchive paleoseismic record of late Pleistocene and Holocene strong earthquakes in Switzerland. Tectonophysics, 400, 153–177.

Behrends, T., Krawczyk-Barsch, E. and Arnold, T. 2012 Implementation of microbial processes in the performance assessment of spent nuclear fuel repositories. Applied Geochemistry 27 453-462. doi:10.1016/j.apgeochem.2011.09.014

Bell, M. and Walker, M. J. C. 2005. Late Quaternary Environmetal Change: Physical and Human Perspectives. (Pearson Education: Harlow).

Berendsen, H.J.A. 1998. Birds-eye view of the Rhine–Meuse delta (The Netherlands). Journal of Coastal Research, 14, 740–752.

Berger, A. and Loutre, M. F. 1996. Modelling the climate response to astronomical and CO_2 forcings. Comptes rendus de l'Académie des sciences. Série 2. Sciences de la terre et des planets. 323, 1-16.

Berger, A. and Loutre, M F. 2003. Climate 400,000 years ago, a key to the future? Geophysical monograph. 137, 17-26

Berger A. and Loutre M. F. 2002. An Exceptionally Long Interglacial Ahead? Science, 297, 1287-1288.

Berger, A., Loutre, M. F., and Gallée, H. 1996. Sensitivity of the LLN 2-D climate model to the astronomical and CO_2 forcings (from 200 kyr BP to 130 kyr AP). Université Catholique de Louvain, Institute d'Astronomie et de Géophysique. Scientific Report 1996/1.

Bertotti, G., Matenco, L. and Cloetingh, S., 2003. Vertical motions in and around the Southeast Carpathians: controls by lithosphere memory and stress field. Terra Nova, 15, 349–359.

Bevan, T. G. and Hancock, P. L. 1986. A late Cenozoic regional mesofracture system in southern England and northern France. Journal Geological Society London, 143, 355-362.

Beylich, A.A., Schimdt, K.H., Neuvonen, S., Forbrich, I. and Schildt, A. 2006. Solute fluxes in the Kidijoki catchment, Subarctic Fuinnish Lapland. Geomorphologie – Relief Processus Environment, 3, 205-212.

Bierman, P.R. 1994. Using in situ cosmogenic isotopes to estiomate rates of landscape evolution: A review from the geomorphic perspective. J. Geophys. Res., 99, 13885-13896.

Bierman, P.R. and Turner J. 1995. Be10 and Al26 evidence for exceptionally low rates of Australian bedrtock erosion and the likely exiostence of pre-Pleistocene landscapes. Quaternary Research, 44, 367-377.

Bills, BG; Currey, DR; and Marshall, GA; 1994; Viscosity estimates for the crust and upper mantle from patterns of lacustrine shoreline deformation in the eastern Great Basin, J. Geophys. Res., 99, 22,059–22,086.

Binnie S E, Phillips W M, Summerfield M A, Fifield L K, Spotila J A, 2008. Patterns of denudation through time in the San Bernadino Mountains, California: implications for early-stage orogenesis. Earth and Planetary Science Letters, 276, pp 62–72

BIOCLIM 2001. Deliverable D3: Global climate features over the next million years and recommendation for specific situations to be considered.

BIOCLIM 2003a. Deliverable D7: Continuous climate evolution scenarios over Western Europe (1000 km scale).

BIOCLIM 2003b. Deliverable D6a: Regional climatic characteristics for the European sites at specific times: the dynamical downscaling.

BIOCLIM 2003c. Deliverable 8a: Development of the rule-based downscaling methodlogy for BIOCLIM Workpackage 3.

BIOCLIM 2003d. Deliverable 8b: Development of the physical/statistical downscaling methodology and application to climate model CLIMBER Workpackage 3.

BIOCLIM website http://www.andra.fr/bioclim/documentation.htm (accessed 7-11-2012)

Biot, M.A., 1941. General theory of three-dimensional consolidation, Journal of Applied Physics 12: 155–164.

Biot, M.A., 1956a. Theory of propagation of elastic waves in fluid-saturated porous solid. I: Low-frequency range. Journal of the Acoustical society of America, 28: 168–178.

Biot, M.A., 1956b. Theory of propagation of elastic waves in fluid-saturated porous solid. I: Higher frequency range. Journal of the Acoustical Society of America 28: 179–191.

Birgersson, M., Karnland, Ol., Nilson, U. 2008. Freezing in bentonite – a thermodynamic approach. Physics and Chemistry of the Earth, 33, pp.527-530.

Birgersson, M., Karnland, O., Nilson, U. 2010. Swelling pressure in compacted bentonite below 0° C. Clays in Natural and Engineered Barriers for Radioactive Waste Confinement, 4th International Meeting. March 2010, Nantes, France.

Bjørlykke, K., Nedkvitne, T., Ramm, M. and Girish, C. 1992. Diagenetic processes in the Brent Group (Middle Jurassic) reservoirs of the North Sea, an overview. In: Morton, A.C., Haszeldine, R.S., Giles, M.R. and Brown, S. (editors). Geology of the Brent Group, Geological Society of London Special Publication, No.61, 263-287.

Blomqvist, R., Kaija, J., Lampinen, P., Paananen, M., Ruskeeniemi, T., Korkealaakso, J., Pitkanen, P., Ludvigson, J.E., Smellie, J., Koskinen, L., Fiona, E., Turrero, M.J., Galarza, G., Jakobsson, K., Laaksoharju, M., Casanova, J., Grundfelt, B. and Hernan, P. (1998) The Palmottu Natural Analogue Project - Hydrogeological Evaluation of the Palmottu Natural Analogue Study Site. European Commission, Nuclear Science and Technology Series, EUR 18202, 96 pp.+ Appendix.

Bluck, B.J., Gibbons, W. and Ingham, J.K., 1992. Terranes. In: Cope, J.C., Ingham, J.K., Rawson, P.F. (Eds.), Atlas of Palaeogeography and Lithofacies: Geol. Soc. (London) Memoirs, 13, 1–4.

Blundell, D. J. 2002. Cenozoic inversion and uplift of southern Britain. In: Doré , A. G., Cartwright, J. A., Stoker, M. S., Turner, J. P. and White, N. (eds) Exhumation of the North Atlantic Margin: Timing, Mechanisms and Implications for Petroleum Exploration. Geological Society, London, Special Publications, 196, 85–101.

Boldreel, L.O. and Andersen, M.S. 1995. The relationship between the distribution of Tertiary sediments, tectonic processes and deep-water circulation around the Faeroe Islands. In: Scrutton, R.A., Shimmield, G.B., Stoker, M.S. and Tudhope, A.W. (eds.), The Tectonics, Sedimentation and Palaeoceanography of the North Atlantic Region. Geological Society, London, Special Publication, 90, 145–158.

Boles, J.R. and Franks, S.G. 1979. Clay diagenesis in Wilcox sandstones of southwest Texas: implications of smectite diagenesis on sandstone cementation. Journal of Sedimentary Petrology, 49, 55-70.

Bond, G, and Lotti, R. 1995. Iceberg discharges into the North Atlantic on millennial time scales during the last glaciation. Science, Vol. 267, 1005–1010.

Bond, G.C. and Kominz, M.A. 1988. Evolution of thought on passive continental margins from the origin of geosynclinal theory (~1860) to the present. Geological Society of America Bulletin, 100 (12), 1909–1933.

Bond, G., Showers, W., Cheseby, M., Lotti, R., Almasi, P., de Menocal, P., Priore, P., Cullen, H., Hajdas, I. and Bonani, G. 1997. A Pervasive Millennial-Scale Cycle in North Atlantic Holocene and Glacial Climates. Science, 278, 1257-1266.

Bond, G. C., Showers, W., Elliot, M., Evans, M., Lotti, R., Hajdas, I., Bonani, G., Johnson, S. 1999. The North Atlantic's 1–2 kyr climate rhythm: relation to Heinrich events, Dansgaard/Oeschger cycles and the little ice age. 59–76 In Clark, P.U., Webb, R.S., Keigwin, L.D. (eds). Mechanisms of Global Change at Millennial Time Scales. Geophysical Monograph. American Geophysical Union, Washington DC.

Bondevik, S. Svendsen, J.I. Johnsen, G. Mangerud J. and Kaland, P.E. 1997. The Storegga tsunami along the Norwegian coast, its age and run-up, Boreas, 26, 29–53.

Bonnet, S., Guillocheau, F. and Brun, J.-P. 1998. Relative uplift measured using river incision: the case of the Armorican basement (France). Earth and Planetary Sciences, 327, 245–251.

Bonnet, S., Guillocheau, F., Brun, J.-P. and Van Den Driessche, J. 2000. Large-scale relief development related to Quaternary tectonic uplift of a Proterozoic–Paleozoic basement: the Armorican Massif, NW France. Journal of Geophysical Research, 105, 19273–19288.

Borlase, W., 1755. Letter to the Rev Charles Lytteton. Philosophical Transactions of the Royal Society of London, 49: 373-378.

Boston, C. M. 2012. A Lateglacial Plateau Icefield in the Monadliath Mountains, Scotland: reconstruction, dynamics and palaeoclimatic implications (Unpublished PhD. Thesis, Queen Mary, University of London).

Bott, M.H.P. and Bott, J.D.J., 2004. The Cenozoic uplift and earthquake belt of mainland Britain as a response to an underlying hot, low-density upper mantle. J. Geol. Soc. (London) 161, 19–29.

Boulton, G. S. 1992. Quaternary. 413-444, in Duff, P. Mcl. D. and Smith, (editors). The Geology of England and Wales. (1st edition) (The Geological Society: London).

Boulton, G. and Hagdorn, M. 2006. Glaciology of the British Isles Ice Sheet during the last glacial cycle: form, flow, streams and lobes, Quaternary Science Reviews, 25, 3359-3390.

Boulton, G.S. and. Payne, A. 1993. Simulation of the European ice sheet through the last glacial cycle and prediction of future glaciation. SKB Tech. Rep. 93-14.

Boulton, G. S., Jones, A. S., Clayton, K. M., and Kenning, M. J. 1977. A British ice-sheet model and patterns of glacial erosion and deposition in Britain 231–246 in British Quaternary studies: recent advances. Shotton, F W (editor). (Oxford: Clarendon Press.)

Boulton, G. S., Smith, G. D., Jones, A. S., and Newsome, J. 1985.Glacial geology and glaciology of the last mid-latitude ice sheets. Journal of the Geological Society of London, 142, 447–474.

Boulton, G.S., Peacock, J.D. and Sutherland, D.G., 1991. Quaternary. In: CRAIG, G.Y. (Ed.), Geology of Scotland. The Geological Society, London, 503-543.

Boulton G S, Caban P E and van Gijssel K, 1995. Groundwater flow beneath ice sheets: part 1 – large scale patterns. Quaternary Science Reviews, 14, 545-562.

Boulton, G., Gustafason, G., Schelkes, K., Casanova, J. and Morén, L. 2001. Palaeohydrogeology and geoforcasting for performamce assessment in geosphere repositories for radioactive waste disposal (PAGEPA). Report, European Commission, Nuclear Science and Technology, EUR 19784, D-G for Research, European Commission, Brussels.

Boulton, G.S., Kautsky, U., Morén, L. and Wallroth, T., 2001. Impact of long-term climate change on a deep geological repository for spent nuclear fuel. SKB Technical Report No. TR-99-05.

Boulton, G. S., Peacock, J. D. and Sutherland, D. G. 2002. Quaternary. 409-430, in Trewin, N. H. (editor). The Geology of Scotland. (2nd edition) (The Geological Society: London).

Bowen, D. Q. (editor). 1999a. A revised correlation of Quaternary deposits in the British Isles. Geological Society Special Report, No. 23.

Bowen, D. Q. 1999b. Only four major 100-ka glaciations during the Brunhes Chron? International Journal of Earth Sciences, 8, 276-284

Bowen, D. Q. 1999c. Wales. 80-90 in A revised correlation of Quaternary deposits in the British Isles. Bowen, D. Q. (editor). Geological Society Special Report, No. 23

Bowen, D.Q., Phillips, F.M. McCabe, A.M., Knutz, P.C. and Sykes, G.A. 2002. New data for the Last Glacial Maximum in Great Britain and Ireland, Quaternary Science Reviews, 21, 89-101.

Bowen, D. Q., Rose, J., McCabe, A. M. and Sutherland, D. G. 1986. Correlation of Quaternary glaciations in England, Ireland, Scotland and Wales. 299–340 in Quaternary Glaciations in the Northern Hemisphere. Sibrava, V., Bowen, D. Q. and Richmond, G. M. (editors). Quaternary Science Reviews, 5.

Brace,W. F. and Byerlee, J.D., 1966. Stick slip as a mechanism for earthquakes, Science, 153, 990–992.

Brace, W.F. and Kohlstedt, D.L. 1980. Limits on lithostatic stress imposed by laboratory experiments. Journal of Geophysical Research, 85, 6248–6252.

Brace. W.F., Paulding, J., and Scholz, C., 1966. Dilatancy in the fracture of crystalline rocks. Journal of the Geophysical Research, 71, 3939–3953.

Braconnot, P., Otto-Bliesner, B., Harrison, S., Joussaume, S., Peterchmitt, J.-Y., Abe-Ouchi, A., Crucifix, M., Driesschaert, E., Fichefet, Th., Hewitt, C. D., Kageyama, M., Kitoh, A., Laıne A., Loutre, M.-F., Marti, O., Merkel, U., Ramstein, G., Valdes, P., Weber, Yu, Y. and Zhao, Y. 2007. Results of PMIP2 coupled simulations of the Mid-Holocene and Last Glacial Maximum – Part 1: experiments and large-scale features. Clim. Past, 3, 261–277.

Bradley, S. L., Milne, G. A., Shennan, I. and Edwards, R. 2011. An improved glacial isostatic adjustment model for the British Isles. J. Quaternary Sci., 26: 541–552

Bradley, S. L., Milne, G. A., Teferle, F. N., Bingley, R. M. and Orliac, E. J. 2009. Glacial isostatic adjustment of the British Isles: new constraints from GPS measurements of crustal motion. Geophysical Journal International, 178, 14–22.

Bradwell, T, Stoker, M S, Golledge, N R, Wilson, C K, Merritt, J W, Long, D, Everest, J D, Hestvik O B, Stevenson, A G, Hubbard, A L. Finlayson, A G and. Mathers, H E. 2008. The northern sector of the last British Ice Sheet: Maximum extent and demise, Earth-Science Reviews, 88, 207-226.

Brand, D., Booth, S.J. and Rose, J. 2002. Late Devensian glaciations, ice damned lake and river diversion, Stiffkey, north Norfolk, England. Proceedings of the Yorkshire Geological Society, 54, 35-46.

Brekke, H. 2000. The tectonic evolution of the Norwegian Sea Continental Margin with emphasis on the Vøring and Møre Basins. In: Nøttvedt, A. (ed.), Dynamics of the Norwegian Margin. Geological Society, London, Special Publications, 167, 327–378.

Brereton, N.R. and Muller, B. 1991. European stress: contributions from borehole breakouts. In: Whitmarsh, R.B., Bott, M.H.P., Fairhead, J.D. and Kusznir, N.J. (eds.). Royal Society Discussion Meeting - Origin and distribution of tectonic stress in the lithosphere. Philosophical Transactions of the Royal Society of London, A337 (1645), 165-179.

Bretagnon, P. and Francou, G. 1988. Planetary theories in rectangular and spherical variables. VSOP87 solutions. Astronomy and Astrophysics, 202: 309–315.

Bridgland, D.R. 1994. The Pleistocene of the Thames. In Quaternary of the Thames. Geological Conservation Review Series, No. 7, Chapman and Hall, London.

Bridgland, D.R. 2010. The record from British Quaternary river systems within the context of global fluvial archives. Journal of Quatenary Science, 25(4) 433-446.

Broecker, W. S. 1994. Massive iceberg discharges as triggers for global climate change. Nature, 372, 421–424.

Broecker, W. S. and Denton, G. H. 1990. What drives glacial cycles? Scientific American, 262, 43–50.

Brodsky, E.E., Roeloffs, E., Woodcock, D., Gall, I., and Manga, M., 2003. A mechanism for sustained ground water pressure changes induced by distant earthquakes. Journal of Geophysical Research, 108

Brooks, A. J., Bradley, S. L., Edwards, R. J., Milne, G. A., Horton, B. and Shennan, I. 2008. Postglacial relative sea-level observations from Ireland and their role in glacial rebound modelling. J. Quaternary Sci., 23. 175–192.

Brown, E.T., Stallard, R.F., Larsen, M.C., Larsen, G.M. Raisbeck and Yiou, F. 1995. Denudation rates determined from the accumulation of in-situ produced 10Be in the Luquillo Experimental Forest, Peurto Rico. Earth Planet. Sci. Lett. 129, 193-202.

Buckley, D.K., Hinsby, L.K. and Manzano, M., 2001. Application of geophysical borehole logging techniques to examine coastal aquifer palaeohydrogeology. *From:* EDMUNDS, W. M. and MILNE, C. J. (eds). 2001. *Palaeowaters in Coastal Europe: evolution of groundwater since the late Pleistocene.* Geological Society, London, Special Publications, 189, 251-270.

Burley, S.D. 1984. Patterns of Diagenesis in the Sherwood Sandstone Group (Triassic), United Kingdom. Clay Minerals, 119, 403-440.

Burley, S.D. and Kantorowicz, J.D 1986. Thin section and S.E.M. criteria for the recognition of cement-dissolution porosity in sandstones. Sedimentology, 33, 587-604.

Burke, H., Phillips, E., Lee, J.R. and Wilkinson, I.P. 2009. Imbricate thrust stack model for the formation of glaciotectonic rafts: an example from the middle Pleistocene of north Norfolk, U.K. Boreas, 38, 620-637.

Burov, E.B. and Watts, A.B. 2006. The long-term strength of the continental lithosphere: 'jelly sandwich' or 'créme brulée'? GSA Today, 16, 4–10.

Busby, J; 2010 Geothermal Prospects in the United Kingdom; Proceedings World Geothermal Congress 2010, Bali, Indonesia, 25-29 April 2010 pp 1-7

Byrd, J.O.D., Smith, R.B. and Geissman, J.W. 1994. The Teton fault, Wyoming: Topographic signature, neotectonics, and mechanisms of deformation. Journal of Geophysical Research, 99, 20095–20122.

Calais, E., Freed, A.M., Van Arsdale, R. & Stein, S. 2010. Triggering of New Madrid seismicity by late-Pleistocene erosion. Nature, 466, 608-612.

Callum RF and Stewart, IS; 2000; Postglacial tectonics of the Scottish glacio-isostatic uplift; Quaternary Science Reviews, 19, pp. 1469–1493.

Carr, S. J., Holmes, R., van der Meer, J. J. M. and Rose, J. 2006. The Last Glacial Maximum in the North Sea Basin: micromorphological evidence of extensive glaciation. J. Quaternary Sci., 21, 131–153.

Catt, J. A., Gibbard, P. L., Lowe, J. J., McCarrol, D., Scource, J. D., Walker, M. J. C. and Wymer, J. J. 2006. Quaternary: ice sheets and their legacy. 429-467, in Brenchley, P. J. and Rawson, P. F. (editors). The Geology of England and Wales. (2nd edition) (The Geological Society: London).

Cedercreutz, J. 2004. Future climate scenarios for Olkiloto with emphais on permafrost. Posiva Report 2004-2006.

Chalmers, J.A., 2000. Offshore evidence for Neogene uplift in central West Greenland. Global Planetary Change, 24, 311–318.

Chapman, N.A., Tsuchi, H. and Kitayama, K. 2009. 1. Tectonic events and nuclear facilities. In: Connor, C.B., Chapman, N.A. and Connor, L.J. (eds), Volcanic and tectonic hazard assessment for nuclear facilities. Cambridge University Press, 1-23.

Charlesworth, C. K. 1957. The Quaternary era: With special reference to its glaciation. (Edward Arnold: London).

Chen, W.-P. and Molnar, P. 1983. Focal depths of intracontinental and intraplate earthquakes and their implications for the thermal and mechanical properties of the lithosphere. Journal of Geophysical Research, 88, 4183–4214.

Christensen, J.H., Hewitson, B., Busuioc, A., Chen, A., Gao, X., Held, I., Jones, R., Kolli, R. K., Kwon, W.-T., Laprise, R., Magaña Rueda, V., Mearns, L., Menéndez, C. G., Räisänen, J., Rinke, A., Sarr, A. and Whetton, P. 2007. Regional Climate Projections. In: Solomon, S. Qin, D., Manning, M., Chen, Z., Marquis, M., Averyt, K. B., Tignor, M. and Miller H. L. (eds.) Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change. Cambridge University Press

Clapperton, C. M. 1997. Greenland ice cores and North Atlantic sediments: implications for the last glaciation in Scotland. 45–58 in Reflections on the ice age in Scotland. Gordon, J E (editor).Glasgow: Scottish Association of Geography Teachers and Scottish Natural Heritage.

Clark, J., McCabe, A. M., Bowen, D. Q. and Clark, P. U. 2012a. Response of the Irish Ice Sheet to abrupt climate change during the last deglaciation, Quaternary Science Reviews, 35, 100-115.

Clark, C. D., Hughes, A. L.C., Greenwood, S. L., Jordan, C. and Sejrup, H. P. 2012b. Pattern and timing of retreat of the last British-Irish Ice Sheet, Quaternary Science Reviews, 44, 112-146.

Clark, P. U., Dyke, A. S., Shakun, J. D., Carlson, A. E., Clark, J., Wohlfarth, B., Mitrovica, J. X., Hostetler S. W. and McCabe, A. M. 2009. The Last Glacial Maximum, Science, 325, 710-714.

Cleaves, E.T., Godfrey, A.E. and Bricker, O.P. 1970. Geochemical balance of a small watershed and its geomorphic implications. Geological Society America Bulletin, 81, 3015-3032.

Cloetingh, S. (Ed.), 2000. Environmental tectonics and climate: NEESDI. Global and Planetary Change, 27, 263pp.

Cloetingh, S., Gradstein, F. M., Kooi, H., Grant, A. C. and Kaminski, M. 1990. Plate reorganization – a cause of rapid late Neogene subsidence and sedimentation around the North Atlantic? Journal of the Geological Society, London, 147, 495–506.

Cloetingh, S; Beekman, F; Zeigler, PA; Van Wees, J-D and Sokoutis, D; 2003; Post-rift compressional reactivation potential of passive margins and extensional basins; In: Johnson, H., Doré, A. G., Gatliff, R. W., Holdsworth, R., Lundin, E. R. and Ritchie, J. D. (eds) The Nature and Origin of Compression in Passive Margins. Geological Society, London, Special Publications, 306, p 27-70.

Cloetingh, S., Ziegler, P.A., Beekman, F., Andriessen, P.A.M., Matenco, L., Bada, G., Garcia-Castellanos, D., Hardebol, N., Dèzes, P. and Sokoutis, D. 2005. Lithospheric memory, state of stress and rheology: neotectonic controls on Europe's intraplate continental topography. Quaternary Science Reviews, 24, 241-304.

Cloos, M. 2009. 2. The nature of tectonic hazards. In: Connor, C.B., Chapman, N.A. and Connor, L.J. (eds), Volcanic and tectonic hazard assessment for nuclear facilities. Cambridge University Press, 24-73.

Cockburn, H.A.P., Seidi, M.A. and Summerfield, M.A. 1999. Quantifying denudation rates on inselbergs in the central Namib Desert using in-situ- produced cosmogenic 10Be and 26Al. Geology, 27(5), 399-402.

Cohen, K.M., Stouthamer, E. and Berendsen, H.J. 2002. Fluvial deposits as a record for Late Quaternary neotectonic activity in the Rhine–Meuse delta, The Netherlands. Netherlands Journal of Geosciences, 81, 389–405.

Collett, T.S. and Bird, K.J. 1988. Freezing point depression at the base of the ice-bearing permafrost on the north slope of Alaska. In: proceedings of the 5th International Conference on permafrost, Trondheim, Norway, August2-5, 50-55.

Coombs, P., Wagner, D., Bateman, K., Harrison, H., Milodowski, A. E., Noy, D. and West, J. M. 2010. The role of biofilms in subsurface transport processes. Q. J. Engineering Geology 43, 131-139. doi:10.1144/1470-9236/08-029.

Condesso De Melo, M.T., Carreira Paquete, P.M.M., and Marques Da Silva, M.A., 2001. Evolution of the Aveiro Cretaceous aquifer (NW Portugal) during the Late Pleistocene and present day: evidence from chemical and isotopic data. *From:* Edmunds, W. M. and Milne, C. J. (eds). 2001. ***Palaeowaters in Coastal Europe: evolution of groundwater since the late Pleistocene.*** Geological Society, London, Special Publications, 189, 139-154

Cooper , A.H., Farrant, A.R. and Price, S.J. 2011. The use of karst geomorphology for planning, hazard avoidance and development in Great Britain. Geomorphology.

Crone, A.J., de Martini, P.M., Machette, M.N., Okumura, K. and Prescott, J.R. 2003. Palaeoseismicity of aseismic Quaternary faults in Australia: Implications for fault behaviour in stable continental regions. Bulletin of the Seismological Society of America, 93, 1913-1934.

Cross, P. and Hodgson, J.M. 1975. New evidence for the glacial diversion of the River Teme near Ludlow, Salop. Proc. Geol. Ass., 86 (3), 313--331.

Davenport, C.A., Ringrose, P.S., 1985. Fault activity and palaeoseismicity during Quaternary time in Scotland - preliminary studies. In Earthquake Engineering in Britain, Thomas Telford Ltd., London, pp. 143-155.

Davenport, C.A., Ringrose, P.S., Becker, A., Hancock, P. and Fenton, C., 1989. Geological investigations of late and post glacial earthquake activity in Scotland. In: Gregersen, S., Basham, P.W. (Eds.), Earthquakes at North Atlantic Passive Margins: Neotectonics and Postglacial Rebound. Academic Publishers, Kluwer Dordrecht, pp. 175-194.

Davies, B. J., Bridgland, D. R., Roberts, D. H., Cofaigh, C. Ó., Pawley, S. M. Candy, I., Demarchi, B.,. Penkman, K. E.H. and Austin, W. E. N. 2009. The age and stratigraphic context of the Easington Raised Beach, County Durham, UK, Proceedings of the Geologists' Association, 120, 183-198.

Davis, M., White, N., Priestly, K., Baptie, B., and Tilmann, F., 2012. Crustal structure of the British Isles and its epeirogenic consequences. Geophys. J. Int., 190(2), 705-725.

Decelles, P.G. and Giles, K.A. 1996. Foreland basin systems. Basin Research, 8, 105-123.

De Geer, G., 1888. Om Skaninaviens nivöförändringar under Quartärperioden. Geologiska Föreningens I Stockholm Förhandlingar, 10, 366-379.

Dehls, J.F., Olesen, O., Olsen, L. and Harald-Blikra, L. 2000. Neotectonic faulting in northern Norway; the Stuoragurra and Nordmannvikdalen postglacial faults. Quaternary Science Reviews, 19, 1447–1460.

DEFRA 2006. Documenting soil erosion rates on agricultural land in England and Wales. Defra report SP0411.

Delisle, G, Caspers, G, and Freund, H, 2003 Permafrost in north-central Europe during the Weichselian: how deep? P 187 – 191 in Permafrost, Phillips, M, Springman, SM & Arenson, LU (eds) 2003 Swets & Zeitlinger, Lisse, ISBN 90 5809 582

De Luca, G., Del Pezzo, E., Di Luccio, F., Margheriti, L., Milana, G. and Scarpa, R., 1998. Site response study in Abruzzo (Central Italy): Underground array versus surface stations. J. Seismology, 2, 223-236.

Dethier, D.P. 1986. Weathering rates and the chemical flux from catchments in the Pacific Northwest, U.S.A. in Cloman, S.M. and Dethier, D.P. (eds), Rates of chemical weathering of rocks and minerals, Academic Press, Orlando, 503-530.

Demoulin, A. 1998. Testing the tectonic significance of some parameters of longitudinal river profiles: the case of the Ardennes (Belgium, NW Europe). Geomorphology, 24, 189–208.

Demoulin, A., Pissart,A. and Zippelt, K. 1995. Neotectonic activity in and around the southwestern Rhenish shield(West Germany): indications of a leveling comparison. Tectonophysics, 249, 203–216.

Department for Environment, Food and Rural Affairs, Department for Business, Enterprise and Regulatory Reform, Welsh Assembly Government, and Department of the Environment Northern Ireland, 2008 Managing Radioactive Waste Safely: A Framework for Implementing Geological Disposal. Cm 7386, ISBN 0101738625, June 2008.Enterprise and Regulatory Reform, Welsh Assembly Government, and Department of the Environment Northern Ireland, 2008 Managing Radioactive Waste Safely: A Framework for Implementing Geological Disposal. Cm 7386, ISBN 0101738625, June 2008.

D'Hondt S, Rutherford S, Spivack AJ. 2002. Metabolic activity of subsurface life in deep-sea sediments. Science 295:2067-2070.

Doré, A. G., Lundin, E. R., Jensen, L. N., Birkeland, Ø., Eliassen, P. E. and Filcher, C. 1999. Principal tectonic events in the evolution of the northwest European Atlantic margin. In: Fleet, A. J. and Boldy, S. A. R. (eds) Petroleum Geology of Northwest Europe: Proceedings of the 5th Conference. The Geological Society, London, 41–61.

Doré, A. G., Cartwright, J. A., Stoker, M. S., Turner, J. P. and White, N. 2002a. Exhumation of the North Atlantic margin: introduction and background. In: Doré, A. G., Cartwright, J. A., Stoker, M. S., Turner, J. P. and White, N. (eds) Exhumation of the North Atlantic Margin: Timing, mechanisms and implications for petroleum exploration. Geological Society, London, Special Publications, 196, 1–12.

Doré, A. G., Corcoran, D. V. and Scotchman, I. C. 2002b. Prediction of the hydrocarbon system in exhumed basins, and application to the NW European margin. In: Doré, A. G., Cartwright, J. A., Stoker, M. S., Turner, J. P. and White, N. (eds) Exhumation of the North Atlantic Margin: Timing, Mechanisms and Implications for Petroleum Exploration. Geological Society, London, Special Publications, 196, 401–429.

Dowdeswell, J.A., Ottesen, D. and Rise, L. 2010. Rates of sediment delivery from the Fennoscandian ice sheet through an ice age. Geology, 38, 3-6.

Dramis, F. and Tondi, E. 2005. Neotectonics. In: Koster, E.A. (ed) The physical Geography of Western Europe, Oxford University Press, 23-37.

Drouet, L., Edwards, N. and Haurie, A. 2006. Coupling climate and economic models in a cost-benefit framework: A convex optimisation approach. Environmental Modelling and Assessment. 11, 101-114.

Edmunds, W.M., 2001. Palaeowaters in European coastal aquifers - the goals and main conclusions of the PALAEAUX project. *From:* Edmunds, W. M. and Milne, C. J. (eds). 2001. Palaeowaters in Coastal Europe: evolution of groundwater since the late Pleistocene. Geological Society, London, Special Publications, 189, 1-16.

Edmunds, W.M., Robins, N.S. and Shand, P., 1998. The saline waters of Llandrindod and Builth, Central Wales. *Journal of the Geological Society, London,* 155, 627–637.

Edmunds, W.M., Buckley, D.K., Darling, W.G., Milne, C.J., Smedley, P.L.and Williams, A.T., 2001a. Palaeowaters in the aquifers of the coastal regions of southern and eastern England. From: Edmunds, W. M. and Milne, C. J. (eds). 2001. Palaeowaters in Coastal Europe: evolution of groundwater since the late Pleistocene. Geological Society, London, Special Publications, 189.

Edmunds, W.M., Hinsby, K., Marlin, C., Condesso De Melo, M.T., Manzano, M., Vaikmae, R. and Travi, Y., 2001b. Evolution of groundwater systems at the European coastline. From: Edmunds, W. M. and Milne, C. J. (eds). 2001. Palaeowaters in Coastal Europe: evolution of groundwater since the late Pleistocene. Geological Society, London, Special Publications, 189, 289-311

Edmunds, W. M and C. J. Milne, 2001. Palaeowaters in Coastal Europe: Evolution of Groundwater Since the Late Pleistocene, edited by, Geological Society Special Publication, 189, 1–16.

Ego, F., Sébrier, M., Carey-Gailhardis, E. and Beate, B. 1996. Do the Billecocha normal faults (Ecuador) reveal extension due to lithospheric body forces in the northern Andes? Tectonophysics, 265, 255-273.

Ehrlich, H. L and Newman, D. K. 2009. Geomicrobiology (5th Edition). CRC Press, Taylor and Francis Group, Florida. 628pp. ISBN-10: 0849379067

Ekman, M. 1991. A concise history of postglacial land uplift research (from its beginning to 1950). Terra Nova, 3, 358-365.

Ekman M. and Mäkinen, J. 1996. Recent postglacial rebound, gravity change and mantle flow in Fennoscandia. Geophysical Journal International, 126, 229-234.

Engelder, T. 1982. Is there a genetic relationship between selected regional joints and contemporary stress within the lithosphere of North America? Tectonics, 1, 161-177.

Engelder, T. 1985. Loading paths to joint propagation during a tectonic cycle: an example from the Appalachian Plateau. Journal of Structural Geology. 7. pp459-476.

Environment Agency 2009 Geological Disposal Facilities on Land for Solid Radioactive Wastes Guidance on Requirements for Authorisation; Environment Agency and Northern Ireland Environment Agency, February 2009.

Evans, CJ; 1987; Crustal stress in the United Kingdom: British Geological Survey, Investigation of the geothermal potential of the U.K. report WJ/GE/87/008; 31p.

Evans, E., Ashley, R., Hall, J., Pennington-Roswell, E. Saul, A., Sayers, P., Thorne, C. and Watkinson, A. 2004. Foresight Future Flooding. Scientific Summary: Volume 1 – Future Risks and Their Drivers. (Office of Science and Technology, London).

Everest, J. D., Bradwell, T., Stoker, M., & Dewey, S. (2012). New age constraints for the maximum extent of the last British–Irish Ice Sheet (NW sector). Journal of Quaternary Science.

Eyles, N., 1993. Earth's glacial record and its tectonic setting. Earth Science Reviews, 35, 1-248.

Firth, C. R. 1989. Late Devensian shorelines and ice limits in the inner Moray Firth area, Scotland. Boreas, 18, 5-21.

Firth, C.R. and Stewart, I.S., 2000. Postglacial tectonics of the Scottish glacio-isostatic uplift centre", Quat. Sci. Rev., 19, 1469-1493.

Firth, CR, Smith, DE and Cullingford, RA, 1993. Late Devensian and Holocene glacio-isostatic uplift patterns in Scotland. Quaternary Proceedings 3, 1-14.

Fichler, C., Nenriksen, S., Rueslaatten, H. and Hovland, M. 2005. North Sea Quaternary morphology from seismic and magnetic data: indications for gas hydrates during glaciation. Petroeum Geoscience, 11, 331-337.

Fjeldskaar, W. 1994. Viscosity and thickness of the asthenosphere detected from the Fennoscandian uplift. Earth and Planetary Science Letters, 126, 399-410.

Francis, A. J., Gillow, J. B., Dodge, C. J., Harris, R., Beveridge, T. J. and Papenguth, H. W. 2004. Uranium association with halophilic and non-halophilic bacteria and archaea. Radiochim. Acta. 92: 481-488.

Fredén, C. (ed) 1994. National Atlas of Sweden; Geology. Stockholm: SNA Publishing.

Freed, AM; and Bürgmann, R; 2004; Evidence of power-law flow in the Mojave Desert mantle, Nature, 430, 548– 551.

Frey, M. 1987. Low Temperature Metamorphism. Blackie, Glasgow and London, 351pp.

Fukushima, Y., Gariel, J.-C. and Tanaka, R.,1995. Site dependent attenuation relations of seismic motion parameters at depth using borehole data. Bull. Seismol. Soc. Am. 85 (6), 1790-1804.

Gábris, G. 1994. Pleistocene evolution of the Danube in the Carpathian basin. Terra Nova, 6, 495–501.

Gallée, H., van Ypersele, J.P., Fichefet, T., Tricot, C., and Berger, A. 1991. Simulation of the last glacial cycle by a coupled, sectorially averaged climate-ice sheet model. 1. The Climate Model. Journal of Geophysical Research, 96, 139-161.

Gallée, H., van Ypersele, J.P., Fichefet, T., Tricot, C., and Berger, A. 1992. Simulation of the last glacial cycle by a coupled, sectorially averaged climate-ice sheet model. 2. Response to Insolation and CO_2 Variations. Journal of Geophysical Research, 97, 713-740.

Garnett, E. R., Andrews, J. E., Preece, R. C. and Dennis, P. F. 2004. Climatic change recorded by stable isotopes and trace elements in a British Holocene tufa. J. Quaternary Sci., 19, 251–262.

Geiger, A., Kahle, H.-G. and Gubler, E. 1986. Recent crustal movements in the Alpine-Mediterranean region analysed in the Swiss Alps. Tectonophysics, 130, 289-298.

Geirsdóttir Á, Miller G H, Andrews J T, 2007. Glaciation, erosion and landscape evolution of Iceland. Journal of Geodynamics, 43, pp 170–186.

Gheorghiu, D.M., Fabel, D., Hansom, J.D. and Xu, S. 2012. Lateglacial surface exposure dating in the Monadhliath Mountains, Central Highlands, Scotland, Quaternary Science Reviews, 41, 132-146.

Gibbard, P. L. and Clark, C. D. 2011. Pleistocene glaciation limits in Great Britain. 75-93. In:. Elhers, J., Gibbard, P. L. and Hughes, P.D. (eds), Developments in Quaternary Science, 15: Quaternary Glaciations - Extent and Chronology: A Closer Look. Elsevier.

Gibbard, P. L., Pasanen, A. H., West, R. G., Lunkka, J. P., Borehham, S., Cohen, K. M. and Rolfe, C. 2009. Late Middle Pleistocene glaciations in East Anglia, England. Boreas, 38, 504-528.

Gibbard, P. L., Rawson, J. E. and Smith, A. G. 2006. The Future. 469-475, in Brenchley, P. J. and Rawson, P. F. (editors). The Geology of England and Wales. (2nd edition) (The Geological Society: London).

Glynn, P., Voss, C. and Provost, A.M., 1999. Deep penetration of oxygenated meltwaters from warm based ice-sheets into the Fennoscandian Shield. In: Use of Hydrogeochemical Information in Testing Groundwater Flow Models, Proc. NEA-SEDE Workshop, Borgholm, Sweden, September 1997. OECD Nuclear Energy Agency, Paris, 201-242

Godwin, H. 1975. The history of British flora: a factual basis for phytogeography. (2nd Edition). (Cambridge University Press).

Goodess, C.M., Watkins, S.J., Palutikof, J.P. and Thorne, M.C. 2004. The construction of global eustatic sea-level scenarios for the next 150,000 years. Climatic Research Unit Research Paper Number 3 (Second Series) Version 4, October 2004

Golledge, N. R., Hubbard, A. and Sugden, D. E. 2008. High-resolution numerical simulation of Younger Dryas glaciation in Scotland. Quaternary Science Reviews, 27, 888-904.

Gourmelen, N. and Amelung, F. 2005. Postseismic mantle relaxation in the Central Nevada Seismic Belt. Science, 310, 1473–1476.

Grassmann, S.; Cramer, B.; Delisle, G.; Hantschel, T.; Messner, J.; Winsemann, J. 2010; pT-effects of Pleistocene glacial periods on permafrost, gas hydrate stability zones and reservoir of the Mittelplate oil field, northern Germany. Marine and Petroleum Geology vol. 27 issue 1, 2010. p. 298-306.

Grollimund, B. and Zoback, M.D. 2001. Did deglaciation trigger intraplate seismicity in the New Madrid seismic zone? Geology, 29, 175-178.

Grünthal, G., Bosse, C., Musson, R.M.W., Gariel, J.-C., Crook, T.D., Verbeiren, R., Camelbeek, T., Mayer-Rosa, D. and Lenhardt, W., 1996. Joint seismic hazard assessment for the central and western part of GSHAP Region 3 (Central and Northwest Europe). 339-342 in Seismology in Europe. Thorkelsson, B. (editor). (Reykjavik: Icelandic Met. Office.)

Guimerà, J., Duro, L., Jordana, S. and Bruno, J., 1999. Effects of ice melting and redox front migration in fractured rocks of low permeability. Swedish Nuclear Fuel and Waste Management Company (SKB), Technical Report, TR-99-19, Stockholm.

Gunnell, Y., 1998, Present, past and potential denudation rates: Is there a link? Tentative evidence from fission-track data, river sediment loads and terrain analysis in the South Indian shield: Geomorphology, 25, 135–153.

Gupta, S.H., Underhill, J.R., Sharp, I.R. and Gawthorpe, R.L. 1999. Role of fault interactions in controlling syn-rift dispersal patterns: Miocene, Abu Alaqa Group, Suez Rift, Sinai, Egypt. Basin Research, 11, 167.Gupta, H., Chakrapani, G., Selvaraj, K., and Kao, S-J. 2011. The fluvial geochemistry, contributions of silicate, carbonate and saline-alkaline components to chemical weathering flux and controlling parameters: Narmada River (Deccan Traps), India. Geochimica et Cosmochimica Acta, 75, 800-824.

Gutenberg, B. and Richter, C.F., 1944. Frequency of earthquakes in California, Bulletin of the Seismological Society of America, 34, 4, 185-188.

Hagdorn, M. K. M. 2003. Reconstruction of the Past and Forecast of the Future European and British Ice Sheets and Associated Sea–Level Change. (Unpublished PhD. Thesis, University of Edinburgh).

Hall, A.M. and Glasser, N.F. 2003. Reconstructing the basal thermal regime of an ice stream in a landscape of selective linear erosion: Glen Avon, Caingorm Mountains, Scotland. Boreas, 32, 191-207.

Hall, A. M., Auton, C. A., Mcl. Michie, U., Pearson, S. G. and Riding, J. B. 2011. Switching flow patterns within the last ice sheet in northern Scotland. Scottish Journal of Geology, 47, 157-167.

Hallbeck, L. and Pedersen, K. 2008. Characterisation of microbial processes in deep aquifers of the Fennoscandian Shield. Applied Geochemistry. 23: 1796-1819.

Hampel, A. and Hetzel, R. 2006. Response of normal faults to glacial–interglacial fluctuations of ice and water masses on Earth's surface. Journal of Geophysical Research, 111, B046406.

Hampel, A., Hetzel, R. and Densmore, A.L. 2007. Postglacial slip rate increase on the Teton normal fault, northern Basin and Range Province, caused by melting of the Yellowstone ice cap and deglaciation of the Teton Range? Geology, 35, 1107–1110.

Hampel, A., Karow, T., Maniatis, G. and Hetzel, R. 2010. Slip rate variations on faults during glacial loading and post-glacial unloading: implications for the viscosity structure of the lithosphere. Journal of the Geological Society, London, 167, 385–399.

Hancock, P.L. 1985. Brittle microtectonics: principles and practise. Journal Structural Geology, 7, 437-457.

Hancock, 1991. Determining Contemporary Stress Directions from Neotectonic Joint Systems [and Discussion]. Philosophical Transactions Royal Society London, A337, 29-40.

Hancock, P.L. and Engelder, T. 1989. Neotectonic joints. Bulletin Geological Society America, 101, 1197-1208.

Hancock, P.L. and Williams, G.D. 1986. Neotectonics. Journal of the Geological Society, London, 143, 325-326.

Harrar, W. G., A. T. Williams, J. A. Barker and M. Van Camp, 2001. Modelling scenarios for the emplacement of palaeowaters in aquifer systems. *From:* EDMUNDS, W. M. and MILNE, C. J. (eds). 2001. *Palaeowaters in Coastal Europe: evolution of groundwater since the late Pleistocene Geological Society, London, Special Publications* 2001, 189, 213-229.

Harris, C. 2002. Middle and late Quaternary permafrost and periglacial environments in the UK: A review of geological evidence, Cardiff University report to United Kingdom Nirex Limited KSGE050

Harrison, H., Wagner, D., Yoshikawa, H., West, J. M. Milodowski, A.E., Sasaki, Y., Turner, G., Lacinska, A., Holyoake, S., Harrington, J., Noy, D., Coombs, P., Bateman, K. and Aoki, K. 2011, Microbiological influences on fracture surfaces of intact mudstone and the implications for geological disposal of radioactive waste. Mineralogical Magazine, 75, 2449-2466. doi: 10.1180/minmag.2011.075.4.2449.

Hartikainen, J., Kouhia, R. and Wallroth, T. 2010. Permafrost simulations at Forsmark using a numerical 2D thermo-hydro-chemical model. SKB TR-09-17.

Hasegawa, H.S. and Basham, P.W. 1989. Spatial correlation between seismicity and postglacial rebound in the Eastern Canada. In: Gregersen, S. and Basham, P.W. (Eds.), Earthquakes at North Atlantic Passive margins: Neotectonics and Postglacial Rebound. Kluwer, Dordrecht, 483-500.

Hashash, Y. M. A., Hook, J. J., Schmidt, B. and Yao, J. I.-C., 2001. Seismic design and analysis of underground structures. Tunnelling and Underground Space Technology 16, 247-293.

Hays, J. D., Imbrie, J. and Shackleton N. J. 1976. Variations in the Earth's Orbit: Pacemaker of the Ice Ages. Science 194 (4270): 1121–1132.

Haug, G. H., Ganopolski, D., Sigman, M. A., Rosell-Mele, A., Swann, G. E. A., Tiedemann, R., Jaccard S. L., Bollmann, J., Maslin, M. A., Leng, M. J. and Eglinton, G. 2004 North Pacific seasonality and the glaciation of North America 2.7 million years ago. Nature, 433, 821-825.

Heijnis, H. and van der Plicht, J. 1992. Uranium/thorium dating of Late Pleistocene peat deposits in NW Europe, uranium/thorium isotope systematics and open-system behaviour of peat layers, Chemical Geology, 94, 161-171,

Heimsath, A.J., Dietrich, W.E., Nishiizumi, K. and Finkel, R.C. 1997. The soil equilibrium function and landscape equilibrium. Nature 388, 358-361.

Heimsath, A.M., Chappell, J.C., Dietrich, W.E., Nishiizumi, K., and Finkel, R.C., 2000. Soil production on a retreating escarpment in southeastern Australia. Geology, 28(9): 787-790.

Heimsath, A.M., Dietrich, W.E., Nishiizumi, K. and Finkel, R.C., 2001. Stochastic processes of soil production and transport: Erosion rates, topographic variation and cosmogenic nuclides in the Oregon Coast Range. Earth Surface Processes and Landforms, 26, 531-552.

Heimsath, A.M., Furbish, D.J. and Dietrich, W.E., 2005. The illusion of diffusion: Field evidence for depth-dependent sediment transport. Geology, 33(12): 949-952.

Heimsath, A.M., Hancock, G.R., and Fink, D., 2009. The 'humped' soil production function: Eroding Arnhem Land, Australia. Earth Surface Processes and Landforms, 34: 1674-1684.

Heinrich, H. 1988. Origin and consequences of cyclic ice rafting in the northeast Atlantic Ocean during the past 130,000 years., Quaternary Research., 29, 142–152.

Hemming, S. R. 2004. Heinrich events: massive Late Pleistocene detritus layers of the North Atlantic and their global imprint. Reviews of Geophysics 42, 1-43.

Hendriks, B.W.H., Andriessen, P.A.M., Dunai, T.J. and Cloetingh, S.A.P.L. 2005. Denudation history of the Northern Scandes range; comparison with the recordfrom southern Norway. Tectonophysics,

Henton, J.A., Raymer, M.R., Ferland, R., Dragert, H., Mazzotti, S. and Forbes, D.L. 2006. Crustal motion and deformation monitoring of the Canadian landmass. Geomatics, 60(2), 173-191. (Canadian Institute of Geomatics).

Hetland, E.A. and Hager, B.H. 2003. Postseismic relaxation across the Central Nevada Seismic Belt. Journal of Geophysical Research, 108, 2394.

Hetzel, R. and Hampel, A. 2005. Slip rate variations on normal faults during glacial–interglacial changes in surface loads. Nature, 435, 81–84.

Hinsby, K., Harrar, W.G., Nyegaard, E., Konradi, E.B., Rasmussen, E.S., Bidstrup, T., Gregersen, U. and Boaretto, E., 2001. The Ribe Formation in western Denmark- Holocene and Pleistocene groundwaters in a coastal Miocene sand aquifer. In: Edmunds, W. M. and Milne, C. J. (eds). 2001. Palaeowaters in Coastal Europe: evolution of groundwater since the late Pleistocene. Geological Society, London, Special Publications, 189, 29-48.

HMSO 1993 Radioactive Substances Act (ISBN 0105412937.) Norwich, HMSO

HMSO 2010 Environmental Permitting (England and Wales) Regulations 2010; E6516 3/2010 106516T 19585

Holbrook, J. and Schumm, S.A. 1999. Geomorphic and sedimentary response of rivers to tectonic deformation: a brief review and critique of a tool for recognizing subtle epeirogenic deformation in modern and ancient settings. Tectonophysics, 305, 287–306.

Holden, P. B., Edwards, N. R., Wolff, E. W., Lang, N. J., Singarayer, J. S., Valdes, P. J. and Stocker, T. F. 2010. Interhemispheric coupling, the West Antarctic Ice Sheet and warm Antarctic interglacials, Clim. Past, 6, 431-443.

Holford, S.P. Green, P.F. Turner, J.P. Williams, G.A., Hillis, R.R., Tappin, D.R. and Duddy, I.R. 2008. Evidence for kilometre-scale Neogene exhumation driven by compressional deformation in the Irish Sea basin system. In: Johnson, H., Doré, A. G., Gatliff, R. W., Holdsworth, R., Lundin, E. R. and Ritchie, J. D. (eds) The Nature and Origin of Compression in Passive Margins. Geological Society, London, Special Publications, 306, 91–119.

Holmes, R. 1997. Quaternary stratigraphy: the offshore record. 72–94 in Reflections on the ice age in Scotland. Gordon, J E (editor). (Glasgow: Scottish Association of Geography Teachers and Scottish Natural Heritage.).

Hough, S.E., Armbruster, J.G., Seeber, L. and Hough, J.F., 2000. On the Modified Mercalli Intensities and Magnitudes of the 1811-1812 New Madrid, Central United States earthquakes, J. Geophys. Res., 105, 23, 839-23, 864.

Hower, J., Eslinger, E.V., Hower, M.E. and Perry, E.A. 1976. Mechanism of burial metamorphism of argillaceous sediments: Mineralogical and chemical evidence. Geological Society of America Bulletin, 87, 725-737.

Hubbard, A. 1999. High-resolution modelling of the advance of the Younger Dryas Ice Sheet and its climate in Scotland, Quaternary Research, 52, 27-43.

Hubbard, A, Bradwell, T, Golledge, N, Hall, A, Patton, H, Sugden, D, Cooper, R and Stoker, M. 2009. Dynamic cycles, ice streams and their impact on the extent, chronology and deglaciation of the British–Irish ice sheet, Quaternary Science Reviews, 28, 758-776.

Hulme, M. 2001.Gradients of temperature and precipitation across the British Isles in the instrumental record, CRU Research paper number 2 (Second Series), Climatic Research Unit, School of Environmental Sciences, University of East Anglia.

Humphreys, P., West, J. M. and Metcalfe, R. 2010. Microbial Effects on Repository Performance. Prepared by Quintessa Ltd. for the Nuclear Decommissioning Authority. Report No. QRS-1378Q-1.

Humlum, O., Instanes, A and Sollid, J.L. 2003. Permafrost in Svalbard: a review of research history, climatic background and engineering challenges. Polar Research, 22(2), 191-215.

Hunt, AG and Malin, PE; 1998; Possible triggering of Heinrich events by ice-load-induced earthquakes; Nature 393, 155-158 doi:10.1038/30218

Hutchinson, J.N. and Thomas-Betts, A. 1990. Extent of permafrost in southern Britain in relation to geothermal flux. Quarterly Journal of Engineering Geology, 23, 387-390.

Huuse, M. and Lykke-Andersen, H. 2000. Over-deepened valleys in the eastern Danish North Sea: morphology and origin. Quaternary Science Reviews, 19, 1233-1253.

Imbrie, J., Hays, J. D., Martinson, D. G., McIntyre, A., Mix, A. C., Morley, J. Paces, J N. G., Prell, W. L. and Shackleton N. J. 1984. The orbital theory of Pleistocene climate: Support from a revised chronology of the marine 18O record, in Milankovitch and Climate, Part I, edited by Berger, A., Imbrie, J., Hays, H., Kukla, G. and Saltzman, B. D. Reidel Publishing, Dordrecht: 1984., p.269-305

Imbrie, J and Imbrie, J Z. 1980. Modelling the climatic response to orbital variations. Science 207, 943-953.

Ingólfsson, Ó. 2004. Quaternary glacial and climate history of Antarctica, in J. Ehlers and P.L. Gibbard (eds.) Quaternary Glaciations - Extent and Chronology, Part III. Elsevier, pp. 3-43.

Itavaara, M., Nyyssonen, M., Kapanen, A., Nousiainen, A., Ahonen, L. and Kukkonen, I. 2011. Characterisation of bacterial diversity to a depth of 1500m in the Outokumpu deep borehole, Fennoscandian Shield. FEMS Microbiological Ecology. 77: 295-309.

Isaksen, K., Holmund, P., Sollid, J.H. and Harris, C. 2001. Three deep Alpine-permafrost boreholes in Svalbard and Scandinavia. Permafrost and Periglacial Processes, 12, 13-25.

Iverson, N.R. 1991. Potential effects of subglacial water-pressure fluctuations on quarrying. Journal of Glaciology, 37, No. 125, 27-36.

James, T.S. and Bent, A.L. 1994. A comparison of eastern North American seismic strain-rates to glacial rebound strain-rates. Geophysical Research Letters, 21, 2127-2130.

James, T.S. and Morgan, W.J. 1990. Horizontal motions due to postglacial rebound. Geophysical Research Letters, 17, 957-960.

James, T.S., Clague, J.J., Wang, K. and Hutchinson, I.I. 2000. Postglacial rebound at the Northern Cascadia Subduction Zone. Quaternary Science Reviews, 19, 15271541.

Jamieson, T.F., 1865. On the history of the last geological changes in Scotland. Geological Society of London Quarterly Journal, 21, 161-203.

Jamieson, T.F. 1882. On the cause of the depression and re-evaluation of the land during the Glacial Period. Geological Magazine, N.S 9, 400-407.

Japsen, P. 1997. Regional Neogene exhumation of Britain and the western North Sea. Journal of the Geological Society, London, 154, 239–247.

Japsen, P. and Bidstrup, T. 1999. Quantification of Neogene erosion in Denmark based on sonic data and basin modelling. Bulletin Geological Society Denmark, 46, 79–99.

Japsen, P. and Chalmers, J.A. 2000. Neogene uplift and tectonics around the North Atlantic: overview. Global and Planetary Change, 24, 165–173.

Jenkins, G. J., Perry, M. C., and Prior, M. J. 2008. The climate of the United Kingdom and recent trends. Met Office Hadley Centre, Exeter, UK.

Jenkins, G. J., Murphy, J.M., Sexton, D. M. H., Lowe, J. A., Jones, P. and Kilsby, C. G. 2009. UK Climate Projections: Briefing Report. Met Office Hadley Centre, Exeter, UK.

Jiménez, M.J., Giardini, D., Grünthal, G. and The Sesame Working Group, 2001. Unified seismic hazard modeling throughout the Mediterranean region. Bolletino di Geofisica Teorica ed Applicata, 42, 3-18.

Johansson, J.M., Davis, J.L., Scherneck, H.G., Milne, G.A., Vermeer, M., Mitrovica, J.X., Bennett, R.A., Jonsson, B., Elgered, G., Elösegui, P., Koivula, H., Poutanen, M., Rönnäng, B.O. and Shapiro, I.I. 2002. Continuous GPS measurements of postglacial adjustment in Fennoscandia 1: geodetic results. Journal Geophysical Research, 107 (B8).

Johnsen, S. J. Dahl-Jensen, D. Gundestrup, N. S., Steffensen, J. P., Clausen, H.B., Miller, H., Masson-Delmotte, V., Sveinsbjörnsdottir, A. E. and White J. 2001. Oxygen isotope and palaeotemperature records from six Greenland ice-core stations: Camp Century, Dye-3, GRIP, GISP2, Renland and NorthGRIP. Journal of Quaternary Science, 16,. 299–307.

Johnson, L, Schneider, J, Zuidema, P, Gribi, P, Mayer, G., and Smith, P. 2002 Project Opalinus Clay - Safety Report, Demonstration of disposal feasibility for spent fuel, vitrified high-level waste and long-lived intermediate-level waste (Entsorgungsnachweis), NAGRA, Wettingen, Switzerland. Nagra Technical Report 02-05

Johnston, A.C. 1987. Suppression of earthquakes by large continental ice sheets. Nature, 330, 467–469.

Johnston, A.C., 1989. The effects of large ice-sheets on earthquake genesis. In: Gregersen, S. and Basham, P.W. (Eds.), Earthquakes at North Atlantic Passive Margins: Neotectonics and Postglacial Rebound. Kluwer, Dordrecht, 581-599.

Johnston, P., Wu, P. and Lambeck, K. 1998. Dependence of horizontal stress magnitude on load dimension in glacial rebound models. Geophys. J. Int. 132. pp41-60. Wiley International.

Johnson, H; Ritchie, JD; Hitchen, K; McInroy, DB; and Kimbell; GS; 2005; Aspects of the Cenozoic deformational history of the Northeast Faroe–Shetland Basin, Wyville–Thomson Ridge and Hatton Bank areas; Geological Society, London, Petroleum Geology Conference series, Vol 6, pp. 993-1007; doi: 10.1144/0060993.

Jones, R. L. and Keen, D. H. 1993. Pleistocene Environments in the British Isles. (Chapman and Hall, London).

Jonsson, S., Segall, P., Pedersen, R., and Bjornsson G., 2003. Post-earthquake ground movements correlated to pore-pressure transients. Nature 424: 179–183.

Kamb, B. and 7 others. 1985. Glacier surge mechanism: 1982- 1983 surge of Variegated Glacier, Alaska. Science, 227 (4686), 469-479.

Kashiwaya, K., Ochiai, S., Sakai, H., and Kawai, T. 2003. Onset of Milankovitch-type climatic oscillations in Lake Baikal sediments at around 4 Ma. Earth and Planetary Science Letters, 213, 185-190.

Kaufmann, G. and Amelung, F. 2000. Reservoir-induced deformation and continental rheology in vicinity of Lake Mead, Nevada. Journal of Geophysical Research, 105, 16341–16358.

Kaufmann, G., Wu, P. and Li, G., 2000. Glacial isostatic adjustment in Fennoscandia for a laterally heterogeneous earth, Geophysical Journal International, 143, 262–273.

Kenner, S.J. and Segall, P. 2000. A mechanical model for intrapate earthquakes: Applications to the New Madrid seismic zone. Science, 289, 2329-2332.

Kerr, R. A. 2012. Experts agree global warming is melting the world rapidly. Science, 338, 1138.

Kjelstad, A., Skogseid, J., Langtangen, H.P., Bjørlykke, K. and Hoeg, K. 2003. Differential loading by prograding sedimentary wedges on continental margins: An arch-forming mechanism. Journal of Geophysical Research, 108, 2036.

Klemman, V. and Wolf, D. 1998. Modelling of stresses in the Fennoscandian lithosphere induced by Pleistocene glaciations. Tectonophysics, 294, 291-303.

Klemperer, S.L. 2006. Crustal flow in Tibet: Geophysical evidence for the physical state of Tibetan lithosphere, and inferred patterns of active flow. In: Law, R.D., Searle, M.P. and Godin, L. (eds) Channel Flow, Ductile Extrusion and Exhumation in Continental Collision Zones. Geological Society, London, Special Publications, 268, 39–70.

Knight, J. 1999. Geological evidence for neotectonic activity during deglaciation of the southern Sperrin Mountains, Northern Ireland. Journal of Quaternary Science, 14, 45-57.

Koizumi, N., Lai, W-C., Kitagawa, Y. and Matsumoto, N., 2004. Comment on ‘‘Coseismic hydrological changes associated with dislocation of the September 21, 1999 Chichi earthquake, Taiwan’’ by Lee M et al. Geophys. Res. Lett., 31; L13603, doi:10.1029/2004GL019897.

Kolderup, C.F. 1913. Norges jordskjælv med særlig hensyn til deres utbredelse I rum og tid. Bergen Museums Aarbog 1913, 8 [with abstract in German].

Koppes, M.N. and Montgomery, D.R. 2009. The relative efficacy of fluvial and glacial erosion over modern to orogenic timescales. Nature Geoscience, 2, September 2009, 644-647.

Kreemer, C., Haines, J., Holt, W.E., Blewitt, G. and Lavalle, D. 2000. On the determination of a global strain rate model. Earth & Planetary Science, Letters, 52, 765-770.

Kroon, D. Austin, W.E.N., Chapman, M.R. and Ganssen, G.M. 1997. Deglacial surface circulation changes in the northeastern Atlantic: temperature and salinity records off NW Scotland on a century scale. Paleoceanography, 12, 755–763

Kujansuo, R., 1964. Nuorista sirroksista Lapissa Summary: Recent faults in Lapland. Geologi, 16, 30-36.

Kukkonen, I.T., Olesen, O., Ask, M.V.S. and the PFDP Working Group. 2010: Postglacial Faults in Fennoscandia: Targets for scientific drilling. Geologiska Föreningen (GFF: Geological Society of Sweden), 132 (Pt. 1, March), 71–81.

Kukla, G. J., Bender, M. L., de Beaulieu, J-L., Bond, G., Broecker, W. S., Cleveringa, P., Gavin, J. E., Herbert, T. D., Imbrie, J., Jouzel, J., Keigwin, L. D., Knudsen, K-L., McManus, J. F., Merkt, J., Muhs, D. R., Müller, H., Poore, R. Z., Porter, S. C., Seret, G., Shackleton, N. J., Turner, C., Tzedakis, P. C. and Winograd, I. J. 2002. Last interglacial climates. Quaternary Research, 58, 2-13.

Kukla, G., Berger, A., Lotti, R., and Brown, J.P. 1981. Orbital signature of interglacials. Nature, 290, 295-300.

Kurz, M.D. 1986. In-situ production of terrestrial cosmogenic helium and some applications to Geochronology. Geochimica Cosmochimica Acta, 50, 2855-2862

Lagarde, J.-L., Baize, S., Amorese, D., Delcaillau, B. and Font, M. 2000. Active tectonics, seismicity and geomorphology with special reference to Normandy (France). Journal of Quaternary Science, 15, 745–758.

Lagerbäck, R., 1979. Neotectonic structures in northern Sweden, Geologiska Foreningens I Stockholm Förhandlinger 100, 271-278.

Lagerbäck, R. 1992. Dating of Late Quaternary faulting in northern Sweden. Journal of the Geological Society, London, 149, 285–291.

Lagerbäck, R. and Sundh, M. 2008. Early Holocen faulting and palaeoseismicity in northern Sweden. SGU Research Paper C836, 80 p.

Lagerklint, I. M. and Wright, J. D. 1999. Late glacial warming prior to Heinrich event 1: the influence of ice rafting and large ice sheets on the timing of initial warming. Geology, 27, 1099–1102.

Lambeck, K. 1991. A model for Devensian and Flandrian glacial rebound and sea-level change in Scotland. 33–62 in Glacial isostacy, sea level and mantle rheology. Sabadini, R., Lambeck, K., and Boschi, E., (editors). (Dordrecht: Kluwer Academic Publishers.)

Lambeck, K. 1993a. Glacial rebound of the British Isles-1: preliminary model results. Geophysical Journal International, 115(3), 941–959.

Lambeck, K. 1993b. Glacial rebound of the British Isles-11. A high resolution, high-precision model. Geophysical Journal International, 115, 960–990.

Lambeck, K. 1995. Late Devensian and Holocene shorelines of the British Isles and North Sea frtom models of glacio-hydro-isostatic rebound. Journal of the Geological Society, London, 152, 437-448

Lambert, A., Courtier, N., Sagawa, G.S., Klopping, F., Winester, D., James, T.M. and Liard, J.O. 2001. New constraints on Laurentide postglacial rebound from absolute gravity measurements. Geophysical Research Letters, 28, 2109-2112.

Larson, K.M. and Van Dam, T. 2000. Measuring postglacial rebound with GPS and absolute gravity. Geophysical Research Letters, 27, 3925-3928.

Lenton, T. 2006. Climate Change to the end of the Millennium. Climatic Change, 76, 7-76.

Lever, DA, Berry, JA, Hoch, AR, Holton, D, Kelly, M and Marshall, TA; 2011 The Treatment of Geosphere Evolution in Post-Closure Performance Assessments; SERCO report TCS/002783/001 Issue 4, 130p

Lidberg, M., Johansson, J.M., Scherneck, H.G. and Davis, J.L. 2007. A improved and extended GPS-derived 3D velocity field of glacial isostatic adjustment (GIA) in Fennoscandia. Journal Geodesy, 113, 1–19.

Lin, L-H, Wang, P-L., Rumble, D., Lippmann-Pipke, J., Boice, E., Pratt, L.M., Sherwood,-Lollar, B., Brodie, E.L., Hazen, T.C., Andersen, G.L., De Santis, T.Z., Moser, D.P., Kershaw, D. and Onstott, T.C. 2006, “Long-term sustainability of a high-energy, low-diversity crustal biome”, Science, 314, pp. 479-482.

Litchfield, N., Ota, Y. and Merritts, D. 2009. 4. Tectonic uplift and subsidence. In: Connor, C.B., Chapman, N.A. and Connor, L.J. (eds), Volcanic and tectonic hazard assessment for nuclear facilities. Cambridge University Press, 116-141.

Lönnqvist, M. and Hökmark, H. 2010. Assessment of potential for glacially induced hydraulic jacking at different depths. SKB report R-09-35.

Loosli, H.H., Aeschbach-Hertig, W., Barbecot, E., Blaser, E., Darling, W.G., Dever, L., Edmunds, W.M., Kipfer, R., Purtschert, R. and Walraevens, K., 2001. Isotopic methods and their hydrogeochemical context in the investigation of palaeowaters. From: Edmunds, W. M. and Milne, C. J. (eds). 2001. Palaeowaters in Coastal Europe: evolution of groundwater since the late Pleistocene. Geological Society, London, Special Publications, 189, 193-212.

Loutre, M. F. 2000. Clues from MIS 11 to predict the future climate - a modelling point of view. Earth and Planetary Science Letters, 212, 213-224.

Loutre, M F. and Berger, A. 2000. Future climatic changes: Are we entering an exceptionally long interglacial? Climatic Change, 46, 61-90.

Loutre, M. F. and Berger, A. 2003. Marine Isotope Stage 11 as an analogue for the present interglacial, Global and Planetary Change, 36, 209-217,

Ludwig, A.O. 1995. The surface of the Holsteinian interglacial sediments as a base level for reconstruction of vertical neotectonic movements in northern Germany. Geosynoptika Geoterm, 3, 31–36.

Lund, B., 2005. Effects of deglaciation on the crustal stress field and implications for endglacial faulting: a parametric study for simple Earth and ice models, SKB Technical Report, TR-05-04. Svensk Kärnbräslehantering AB, Stockholm.

Lund, B. and Näsland, J.O. 2009. 5. Glacial isostatic adjsment: implications for glacially induced faulting and nuclear waste repositories. In: Connor, C.B., Chapman, N.A. and Connor, L.J. (eds), Volcanic and tectonic hazard assessment for nuclear facilities. Cambridge University Press, 142-155.

Lundin, E. and Doré, A.G. 2002. Mid-Cenozoic post-breakup deformation in the ‘passive’ margins bordering the Norwegian–Greenland Sea. Marine and Petroleum Geology, 19, 79–93.

Lundquist, J. and Lagerbäck, R. 1976. The Pärve fault: A late glacial fault in the Precambrian of Swedish Lapland. GFF, 98, 45–51.

Macklin, M. G. and Lewin, J. 2003. River sediments, great floods and centennial-scale Holocene climate change. J. Quaternary Sci., 18, 101–105.

MacAyeal, D. R. 1993. Binge/purge oscillations of the Laurentide Ice Sheet as a cause of the North Atlantic's Heinrich events, Paleoceanography, 8(6), 775–784

Maddy, D. 1999. English Midlands. 28-44 in, Bowen, D Q, (editor).. A revised correlation of Quaternary deposits in the British Isles. Geological Society Special Report, No. 23.

Main, I.D., Irving, D., Musson, R.M.W. and Reading, A., 1999. Constraints on the frequency- magnitude relation and maximum magnitudes in the UK from observed seismicity and glacio-isostatic recovery rates. Geophys. J. Int. 137 (2), 535–550.

Mäkinen, J., Engfeldt, A., Harsson, B.G. et el. 2005. The Fennoscandian land uplift gravity lines 1966-2003. In: Jekeli, C., Bastos, L. and Fernandez, J. (eds) Gravity, Geoid and Space Missions, IAG Symposium, 129, Springer, Berlin, Germany, 328-332.

Manga, M. and C.-Y. Wang, 2007. Earthquake hydrology, in Treatise on Geophysics, G. Schubert editor, 4, 293-320. Elsevier.

Manga, M., 2001. Origin of postseismic streamflow changes inferred from baseflow recession and magnitude-distance relations, Geophys. Res. Lett., 28, 2133–2136.

Mangerud, J., Jansen, E., and Landvik, J.Y. 1996. Late Cenozoic history of the Scandinavian and Barents Sea ice sheets. Global and Planetary Change, 12, 11-26.

Mann, M. E., R. S. Bradley, and M. K. Hughes 1999. Northern hemisphere temperatures during the past millennium: Inferences, uncertainties, and limitations, Geophys. Res. Lett., 26(6), 759–762.

Mann, M. E., Bradley, R. S., and Hughes, M. K. 1998. Global-ScaleTemperature Patterns and Climate Forcing Over the Past Six Centuries, Nature, 392, 779-787,

Mann, M. 2012. The Hockey Stick and the Climate Wars: Dispatches from the Front Lines. Columbia University Press.

Marivoet, J., Van Keer, I., Wemaere, I., Hardy, L., Pitsch, H., Beaucaire, C., Michelot, J-L., Martin, C., Philippot, A.C., Hassanizadeh, M. and Van Weert, F. 2000. A palaeohydrogeological study of the MOL site (PHYMOL project). Report, European Commission, Nuclear Science and Technology, EUR 19146, D-G for Research, European Commission, Brussels.

Marrow, P.C. and Roberts, G. 1984. Source mechanisms of aftershocks of the 1979 ‘Carlisle’ earthquake. Global Seismology Unit Report No. 217, British Geological Survey.

Maslin, M. A., Li, X. S., Loutre, M. F. and Berger, A. 1998. The contribution of orbital forcing to the progressive intensification of Northern Hemisphere glaciation. Quaternary Science Reviews, 17, 411–426.

Maslin, M. A. and Ridgwell, A. J. 2005. Mid-Pleistocene revolution and the 'eccentricity myth'. in Early-middle Pleistocene transitions; the land-ocean evidence. Geological Society Special Publications, 247, 19-34.

Maslin, M., Seidov, D. and Lowe, J. 2001. Synthesis of the nature and causes of rapid climate transitions during the Quaternary. Geophysical Monograph Series, 126, 9–52.

Matsushi Y., Wasaka S., Matsuzaki H., Matsukura Y., 2006. Long-term denudation rates of actively uplifting hillcrests in the Boso Peninsula, Japan, estimated from depth profiling of in situ-produced cosmogenic 10Be and 26Al. Geomorphology, 82, pp 283–294.

McCarroll, D., J. O. Stone, Ballantyne, C. K., Scourse, J. D., Fifield, L. K., Evans, D. J. A. and Hiemstra, J. F. 2010.Exposure-age constraints on the extent, timing and rate of retreat of the last Irish Sea ice stream, Quaternary Science Reviews, 29, 1844-1852.

McCaulay, G.E., Burley, S.D., Fallick, A.E. and Kusznir, N.J. 1994. Palaeohydrodynamic fluid flow regimes during diagenesis of the Brent Group in the Hutton-NW Hutton reservoirs; constraints from oxygen isotope studies of authigenic kaolin and reverse flexural modelling. Clay Minerals, 29, 609-626.

McEwen, T. 1991. The potential significance of permafrost to the behaviour of a deep radioactive waste repository. SKI Report, 91:8.

McGarr, A. and Fletcher, J.B., 2005. Development of ground-motion prediction equations relevant to shallow mining-induced seismicity in the Trail Mountain area, Emery County, Utah. Bull. Seismol. Soc. Am. 95 (1), 31-47.

McKinley, I.G and Alexander, W.R. 2009. 22. Issues for coastal sites. In: Connor, C.B., Chapman, N.A. and Connor, L.J. (eds), Volcanic and tectonic hazard assessment for nuclear facilities. Cambridge University Press, 509-526.

McKinley, I.G and Chapman, N.A. 2009. 24. The impact of subsidence, uplift and erosion on geological repositories for radioactive wastes. In: Connor, C.B., Chapman, N.A. and Connor, L.J. (eds), Volcanic and tectonic hazard assessment for nuclear facilities. Cambridge University Press, 548-565.

Meehl, G.A., T.F. Stocker, W.D. Collins, P. Friedlingstein, A.T. Gaye, J.M. Gregory, A. Kitoh, R. Knutti, J.M. Murphy, A. Noda, S.C.B. Raper, I.G. Watterson, A.J. Weaver and Z.-C. Zhao, 2007. Global Climate Projections. In: Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change [Solomon, S. Qin, D., Manning, M., Chen, Z., Marquis, M., Averyt, K. B., Tignor, M. and Miller H. L. (eds.)]. (Cambridge University Press)

Cainozoic geology and landscape evolution of north-east Scotland Meyer H, Hetzel R, Strauss H, 2008. Erosion rates on different timescales derived from cosmogenic 10Be and river loads: implications for landscape evolution in the Rhenish Massif, Germany. International Journal of Earth Sciences, 99, pp 395–412.

Meyer, W; and Stets, J; 1998; Junge Tektonik im Rheinischen Schiefergebirge und ihre Quantifizierung. Zeitschrift Deutsche Geologische Geselschaft 149: 359-379.

Middlemiss, A. 1983. Instability of chalk cliffs between the South Foreland and Kingsdown, Kent in relation to geological structure. Proceedings of the Geologists' Association of London, 94, 115-22.

Miller, G. H., Geirsdóttir, Á., Zhong, Y., Larsen, D. J., Otto-Bliesner, B. L., Holland, M. M., Bailey, D. A., Refsnider, K. A., Lehman, S. J., Southon, J. R., Anderson, C., Björnsson, H. and Thordarson. T. 2012. Abrupt onset of the Little Ice Age triggered by volcanism and sustained by sea-ice/ocean feedbacks, Geophys. Res. Lett., 39, L02708, doi:10.1029/2011GL050168.

Millin, S. (editor). 2010. UKCP09 sea level change estimates. UKCIP Briefing notes, 7pp.

Milne, G.A., Davis, J.L., Mitrovica, J.X., Scherneck, H.G., Johansson, J.M., Vermeer, M. and Koivula, H. 2001. Space-geodectic constraints on glacial isostatic adjustment in Fennoscandia. Science, 291, 2381–2385.

Milne, G.A., Mitrovica, J.A., Scherneck, H., Davis, J.L., Johansson, J.M., Koivula, H. and Vermeer, M. 2004. Continuous GPS measurements of postglacial adjustment in Fennoscandia: 2. modelling results, Journal Geophysical Research, 109, B02412.

Milne, G.A., Shennan, I., Youngs, B. A. R., Waugh, A. I., Teferle, F. N, Bingley, R. M., Bassett, S. E., Cuthbert-Brown, C. and Bradley, S. L. 2006. Modelling the glacial isostatic adjustment of the UK region. Phil. Trans. R. Soc. A., 364, 931-948.

Milodowski, A.E., Gillespie, M.R., Naden, J., Fortey, N.J., Shepherd, T.J., Pearce, J.M. and Metcalfe, R. 1998. The petrology and paragenesis of fracture mineralization in the Sellafield area, west Cumbria. Proceedings of the Yorkshire Geological Society, 52, 215-241.

Milodowski, A.E., Fortey, N.J., Gillespie, M.R., Pearce, J.M. and Hyslop, E.K. 2002. Synthesis report on the mineralogical characteristics of fractures from the Nirex boreholes in the Sellafield area. British Geological Survey, Technical Report, WG/98/8.

Milodowski, A.E. Tullborg, E.-L., Buil, B., Gómez, P., Turrero, M.-J., Haszeldine, S., England, G., Gillespie, M.R., Torres, T., Ortiz, J.E., Zachariáš, Silar, J., Chvátal, M., Strnad, L., Šebek, O. Bouch, J.E., Chenery, S.R., Chenery, C., Shepherd, T.J. and Mckervey, J.A. 2005. Application of Mineralogical, Petrological and Geochemical tools for Evaluating the Palaeohydrogeological Evolution of the PADAMOT Study Sites. PADAMOT PROJECT - EU FP5 CONTRACT NO. FIKW-CT2001-20129. PADAMOT Project Technical Report, WP2. 228pp + digital appendices.

Mitchell, G. F., Penny, L. F., Shotton, F. W. and West, R. G. 1973. A correlation of Quaternary deposits in the British Isles. Geological Society of London Special Report, No. 4.

Mitrovica, J.X., 1996. Haskell 1935 revisited. J. Geophys. Res., 101, 555– 569.

Mitrovica, J.X. and Peltier, W.R. 1993. Present da secular variations in zonal harmonics of the Earth's geopotential. Journal of Geophysical Research, 98, 4509-4526.

Mitrovica, J.X., Milne, G.A. and Davis, J.L. 2001. Glacial isostatic adjustment on a rotating Earth, Geophysical Journal International, 147, 562–578.

Mohr, P. 1986. Possible Late Pleistocene faulting in Iar (West) Connacht Ireland. Geological Magazine, 123, 545-552.

Montford, A. 2010. The Hockey Stick Illusion: Climategate and the Corruption of Science. (Stacey International).

Montgomery, D. R. and Manga, M., 2003. Streamflow and Water Well Responses to Earthquakes. Science, 300, 2047-2049.

Mörner, N.A. 1977. Past and present uplift in Sweden: glacial isostacy, tectonism, and bedrock influence. Geologiska Föreningen I Stockholms Föhandlingar, 99, 48-54.

Mörner, N.-A. 1978. Faulting, fracturing and seismic activity as a function of glacial-isostasy in Fennoscandia. Geology, 6, 41–45.

Mörner, N.-A., 1979. The Fennoscandian uplift and Late Cenozoic Geodynamics: geological evidence. GeoJournal, 3, 287-318.

Mörner, N.-A., 2004. Active faults and paleoseismicity in Fennoscandia, especially Sweden. Primary structures and secondary effects. Tectonophysics 380, 139–157.

Mörner, N.-A. 2005. An interpretation and catalogue of paleoseismicity in Sweden. Tectonophysics, 408, 265–307.

Muir-Wood, R. 1989a. Fifty million years of 'passive margin' deformation in northwest Europe. In: Gregersen, S. and Basham, P.W. (Eds.), Earthquakes at North Atlantic passive margins: neotectonics and Postglacial Rebound. Kluwer Academic Publishers, Dordrecht, 7-36.

Muir-Wood, R., 1989b. Extraordinary deglaciation reverse faulting in northern Fennoscandia, in S. Gregerson and P. Basham (eds.), Earthquakes at North-Atlantic passive margins: Neotectonics and postglacial rebound, Proceedings of the NATO Advanced Research Workshop, Vordingborg, Denmark, 9-13 May 1988, Kluwer, Dordrecht, pp. 141-173.

Muir-Wood R, 1993. A review of the Seismotectonics of Sweden. SKB TR 93-13, Svensk Kärnbränslehantering AB, Stockholm.

Muir-Wood, R., 2000. Deglaciation seismotectonics: A principal influence on intraplate seismogenesis at high latitudes?, Quat. Sci. Rev., 19,1399-1411.

Muir-Wood. R. and King, G.C.P., 1993. Hydrological signatures of earthquake strain. J. Geophys. Res., 98, 22035–22068.

Muller, R.A., and MacDonald, G.J. 1997. Glacial cycles and astronomical forcing. Science, 277, 215-218.

Munier, R. and Fenton, C. 2004. Appendix 3: Review of postglacial faulting. In: Munier, R. and Hökmark, H. (eds) Respect distances. Rationale and Means of Computation. R-report SKB R-04-17. Stockholm, Sweden: Swedish Nuclear Fuel and Waste Management Company.

Musson, R.M.W., 1994. A catalogue of British earthquakes, British Geological Survey, Technical Report WL/94/04.

Musson, R.M.W., 1996. The seismicity of the British Isles. Annali di Geofisica 39, 463–469.

Musson, R.M.W., 1998. The Barrow-in-Furness Earthquake of 15 February 1865: Liquefaction from a Very Small Magnitude Event, Pure and Applied Geophysics, 152(4), 733-745.

Musson, R.M.W. and Sargeant, S.L., 2007. Eurocode 8 seismic hazard zoning maps for the UK, British Geological Survey Technical Report CR/07/125

Musson, RMW, and Winter, PW. 1997 Seismic hazard maps for the U.K.; Natural Hazards; 14; pp141-154.

Nakiboglu, SM and Lambeck, K; 1982; A study of the Earth's response to surface loading with application to Lake Bonneville; Geophysical Journal of the Royal Astronomical Society; Volume 70, Issue 3, pp 577–620.

Nakicenovic, N. and Swart, R. (eds.). 2000. Special Report of IPCC Working Group III - Emissions Scenarios. (Cambridge University Press).

National Research Council of the National Academies (2010). America's Climate Choices: Panel on Advancing the Science of Climate Change, Board on Atmospheric Sciences and Climate, Division on Earth and Life Studies. Sea Level Rise and the Coastal Environment. 243–250 In Advancing the Science of Climate Change. The National Academies Press,Washington, D.C.

New Scientist 2009 http://www.newscientist.com/article/dn11646-climate-myths-the-hockey-stick-graph-has-been-proven-wrong.html (accessed 7-11-2012).

NDA 2010a Geological Disposal: Geosphere status report; Nuclear Decommissioning Authority Report NDA/RWMD/035, 2010.

NDA 2010b Geological Disposal: Biosphere status report. Nuclear Decommissioning Authority Report NDA/RWMD/036, 2010.

NDA, 2010c Geological Disposal – Near-Field Evolution Status Report. NDA Report no. NDA/RWMD/033.

NDA, 2010d. Geological Disposal Generic Post-closure Safety Assessment. NDA Report no. NDA/RWMD/030.

NDA, 2010e. Geological Disposal - Radionuclide behaviour status report. NDA Report no. NDA/RWMD/034.

NEA. 1993. Palaeohydrogeological Methods and their Applications for Radioactive Waste Disposal. Proceedings Symposium Paris, 1992. OECD Nuclear Energy Agency, Paris.

Nicholson, D.T. 2008. Rock control on microweathering of bedrock surfaces in a periglacial environment. Geomorphology, 101, 655-665.

Nishiizumi, K., Khol, C.P., Arnold, J.R., Klein, J., Fink, D. and Middleton, R. 1991. Cosmic ray produced 10Be andc 26Al in Antartic rocks: exposure and erosion history. Earth Planet. Sci. Lett. 104, 440-454.

Nishiizumi, K., Winterer, E.L., Kohl, C.P., Klein, J., Middleton, R., Lal, D., Arnold, J.R. 1993. Role of in-situ cosmogenic nuclides 10Be and 26Al in the study of diverse geomorphic processes. Earth Surface Processes and Landforms, 18, 407-425.

Nishimura, T. and Thatcher, W. 2003. Rheology of the lithosphere inferred from postseismic uplift following the 1959 Hebgen Lake earthquake. Journal of Geophysical Research, 108, 2389.

Niviere, B, and Winter, T; 2000; Pleistocene northwards fold propagation of Jura within the southern Upper Rhine Graben. Seismotectonic implications. In: NEESDI volume. Global and Planetary Change; Vol 27; pp 263–288.

Nuclear Energy Agency; 2005, Stability and buffering capacity of the geosphere for longterm isolation of radioactive waste: application to argillaceous media, Workshop Proceedings, Braunschweig, Germany, 9.11 December 2003. NEA#5303, ISBN: 92-64-00908-6; 244p

Nuclear Energy Agency; 2009, Stability and buffering capacity of the geosphere for longterm isolation of radioactive waste: application to crystalline rock, Workshop Proceedings, Manchester, United Kingdom, 13.15 November 2007. NEA#6362, ISBN: 978-92-64-06056-2, 304p

Ojala, V.J., Kuivamäki, A. and Vuorela, P. 2004. Postglacial deformation of bedrock in Finland. Geological Survey of Finland, Nuclear Waste Disposal Research. Report YST-120, 23 pp.

Olesen, O. 1988. The Stuoragurra Fault; evidence of neotectonics in the Precambrian of Finnmark, northern Norway. Norsk Geologisk Tidsskrift, 68, 107–118.

Ollier, C.D. 1985. Morphotectonics of continental margins with great escarpments. In: Morisawa, M. and Hack, J.T. (eds), Tectonic Geomorphology. Allen and Unwin, Boston; The Binghamton Symposia in Geomorphology, International Series, 15, 3–25.

Olsen, L., Dehls, J.F., Olesen, O. and Rønning, J.S. 1999: Late Quaternary faulting and paleoseismicity in Finnmark, northern Norway. In: Dehls, J.F. and Olesen, O. (eds*)* Neotectonics in Norway. Annual Technical Report 1998. Norges geologiske undersøkelse Report 99.007, 93-102.

Onstott, T.C Mcgown, D.J., Bakermans, Ruskeeniemi, C,R. Ahonen, L. Telling, J. Sofficentino, B., Pfiffner, S.M., Sherwood-Lollar, B., Frape, S., Stotler, R., Johnson, E.J., Vischnivetskaya, T.A., Rothmel, R. and Pratt, L.M. 2009. Microbial communities in subpermafrost saline fracture water at the Lupin Au Mine, Nunavut, Canada. Microbial Ecology, 58, pp 786-807.

Ottemöller, L. and Thomas, C.W., 2007. Highland Boundary Fault Zone: Tectonic implications of the Aberfoyle earthquake sequence of 2003. Tectonophysics 430, 83–95

Ottemöller, L. and Sargeant, S.L., 2010. Ground-motion difference between two moderate-size intraplate earthquakes in the United Kingdom, Bull. Seismol. Soc. Am., 100(4), 1823-1829

Pagli, C., Sigmundsson, F., Lund, B., Sturkell, S., Geirsson, H., Einarsson, P., Árnadóttir, T. and Hreinsdóttir, S. 2007. Glacio-isostatic deformation around the Vatnajökull ice cap, Iceland, induced by recent climate warming: GPS observations and finite element modeling. Journal of Geophysical Research 112, no. B8 B08405..

PALAEUX. 1999. Management of coastal aquifers in Europe: Palaeowaters,natural controls and human influence. Final Report of the Palaeux project ENV-CT95-0156. EC 4th Framework Programme (Climatology and Natural Hazards). British Geological Survey, Report, Hydrogelogy Series, WD/99/36.

Parker, D.E., Legg, T.P. and Folland. C.K. 1992. A new daily Central England Temperature Series, 1772-1991. Int. J. Clim., 12, 317-342

Pascal, C., Stewart, I.S. and Vermeersen, B., 2010. Neotectonics, seismicity and stress in glaciated regions, Journal of the Geological Society, 167, 2, 361-362

Pavich, M. J., 1986, Processes and rates of saprolite production and erosion on a foliated granitic rock of the Virginia piedmont, in Colman, S. M. And Dethier, D. P., eds., Rates of chemical weathering of rocks and minerals: Orlando, Academic Press, Inc., p. 551-590.

Pavich M.J. 1989. Regolith residence time and the concept of surface age of the Piedmont 'peneplain'. Geomorphology, 2, 181-196.

Pearson, D. G, Emeleus, C. H., and Kelley, S. P., 1996. Precise 40Ar/39Ar age for the initiation of Palaeogene volcanism in the Inner Hebrides and its regional significance Journal of the Geological Society of London, 153 (6), 815-818.

Pedersen, K. 2000. Microbial processes in radioactive waste disposal. SKB Technical Report TR-00-04.

Pedersen, K. 2005. Microorganisms and their influence on radionuclide migration in igneous rock environments. J. Nucl. Radiochem. Sci. 6: 11-15.

Peltier, W.R., 1976. Glacial isostatic adjustment ii: the inverse problem, Geophys. J. R. astr. Soc., 46, 669–706.

Peltier, W.R., 1998. A Space geodetic target for mantle viscosity discrimination: horizontal motions induced by glacial isostatic adjustment, Geophysical Research Letters, 25, 543-546.

Peltier,W.R., Shennan, I., Drummond, R. and Horton, B.P. 2002. On the postglacial isostatic adjustment of the British Isles and the shallow viscoelastic structure of the Earth, Geophysical Journal International, 148, 443–475.

Peltzer, G; Rosen, P; Rogez, F and Hudnut, K; 1998; Poroelastic rebound along the Landers 1992 earthquake surface rupture, J. Geophys. Res., 103,30,13 1-30,145.

Petit, J. R., Jouzel, J., Raynaud, D., Barkov, N.I., Barnola, J.-M., Basile, I., Benders, M., Chappellaz, J., Davis, M., Delaygue, G., Delmotte, M., Kotlyakov, V.M.,Legrand, M., Lipenkov, V. Y., Lorius, C., Pepin, L., Ritz, C., Saltzman, E., and Stievenard, M. 1999. Climate and atmospheric history of the past 420 000 years from the Vostok ice core, Antarctica. Nature, 399, 429-436.

Peulvast J-P, Bétard F, Lageat Y, 2009. Long-term landscape evolution and denudation rates in shield and platform areas: a morphostratigraphic approach. Géomorphologie, 2, pp 95–108.

Pfingsten, W., 2001. Indications for self-sealing of a cementitious repository for long and intermediate level waste. Nagra Technical Report, 01-05.

Pharaoh, T., Molyneux, S., Merriman, R., Lee, M. and Verniers, J., 1993. The Caledonides of the Anglo-Brabant Massif reviewed, Geol. Mag. 130, 561–562.

Phillips, E. R., Auton, C. A., Evans, D. J. A. and Benn, D. I. 2003. Drumbeg: glacial sedimentology and tectonostraigraphy, 88-103 in The Quaternary of the Western Highland Boundary: Field Guide., Evans, D. J. A. (editor) (Quaternary Reserach Association: London).

Phillips, W. M., Hall, A. M., Ballantyne, C. K., Binnie, S., Kubik, P. W. and Freeman, S. 2008. Extent of the last ice sheet in northern Scotland tested with cosmogenic 10Be exposure ages. J. Quaternary Sci., 23, 101–107.

Pierce, K.L., Cannon, K.P., Meyer, G.A., Trebesch, M.J. and Watts, R.D. 2002. Post-glacial inflation–deflation cycles, tilting, and faulting in the Yellowstone Caldera based on Yellowstone Lake shorelines. US Geological Survey Open-File Report, 02-0142.

Plafker, G. 1964. Tectonic deformation associated with the 1964 Alaska Earthquake. Science, 148, 1675-1687.

PMIP, Paleoclimate Modelling Intercomparison Project (PMIP) 2000 : proceedings of the third PMIP workshop, Canada, 4-8 October 1999, in WCRP-111, WMO/TD-1007. Braconnot, P. (editor).

Pinot, S., Ramstein, G., Harrison, S. P., Prentice, I. C., Guiot, J., Stute, M. and Joussaume, S. 1999. Tropical paleoclimates at the Last Glacial Maximum: comparison of Paleoclimate Modelling Intercomparison Project (PMIP) simulations and paleodata. Climate Dynamics, 15, 857-874.

Pitkänen, P., Partamies, S. and Luukkonen, A. 2004. Hydrogeochemical interoperation of baseline groundwater conditions at the Olkiluoto site. Posiva Oy. POSIVA 2003-07.

Pollitz, FF; Wicks, C; and Thatcher, W; 2001; Mantle flow beneath a continental strike-slip fault: Postseismic deformation after the 1999 Hector Mine earthquake, Science, 293, 1814– 1818.

Pollitz, F; P. Banerjee, K; Grijalva, B; Nagarajan, and Bürgmann, R; 2008; Effect of 3-D viscoelastic structure on post-seismic relaxation from the 2004 M = 9.2 Sumatra earthquake, Geophys. J. Int. 173 189–204.

Praeg, D., Stoker, M. S., Shannon, P. M., Ceramicola, S., Hjelstun, B., Laberg, J. S. and Mathiesen, A. 2005. Episodic Cenozoic tectonism and the development of the NW European 'passive' continental margin. Marine and Petroleum Geology, 22, 1007–1030.

Preece, R.C. 2001. Molluscan evidence for differentiation of interglacials within the 'Cromerian Complex', Quaternary Science Reviews, 20, 1643-1656.

Preece, R.C. and Parfitt, S. A. 2000. The Cromer Forest-Bed Formation: new thoughts on an old problem.In: Lewis, S.G. Whitman, C.A. and Preece, R. C. (editors) The Quaternary of Norfolk and Suffolk: Field Guide (Quaternary Research Association, London), 1-27.

Preece, R.C., Scourse, J.D., Houghton, S.D., Knudsen, K.L. and Penney, D.N. 1990. The Pleistocene sea-level and neotectonic history of the eastern Solent, southern England. Philosophical Transactions of the Royal Society of London, B328, 425–477.

Price J R, Heitmann N, Hull J, Szymanski D, 2008. Long-term average weathering rates from watershed geochemical mass balance methods: using mineral modal abundance to solve more equation in more unknowns. Chemical Geology, 254, pp 36–51

Pusch, R., Borgesson, L., Knuttson, S., 1990. Origin of silty fracture fillings in crystalline bedrock. Geol. Foren. Stockholm Forh ,112, 209–213.

Quine, T.A. and Walling, D.E. 1991. Rates of soil erosion on arable fields in Britain: quiantiotative data from Caesium-137 measurements. Soil Use and Management, 7, 169-176.

Quinlan, G., 1984. Postglacial rebound and the focal mechanisms of eastern Canadian earthquakes, Ca. J. Earth Sci., 21, 1,018–1,023.

Ramanathan, V., Crutzen, P.J., Kiehl, J.T., and Rosenfeld, D. 2001. Aerosols, climate, and the hydrological cycle. Science, 294, 2119-2124.

Rasmussen, S. O., Andersen, K. K., Svensson, A. M., Steffensen, J. P., Vinther B. M., Clausen, H. B., Siggaard-Andersen, M.-L., Johnsen, S. J., Larsen L. B., Dahl-Jensen, D. Bigler, M., Röthlisberger, R., Fischer H., Goto-Azuma, K. and Hansson, M. E. 2006. A new Greenland ice core chronology for the last glacial termination, J. Geophys. Res., 111, D06102, doi:10.1029/2005JD006079

Raymo, M E. 1994. The initiation of Northern Hemisphere glaciation. Annual Review of Earth and Planetary Sciences, 22, 353–383.

Raymo, M E. 1997. The timing of major climate terminations. Palaeooceanography, 12, 577-585.

Raymo, M. E., and Ruddiman, W. F. 1992. Tectonic forcing of late Cenozoic climate. Nature, 359, 117–122.

Recreo, F., Lomba, L. and Ruiz, C. 2005. Potential climatic and landscape evolution of central Spain (Toledo area) over the next 200,000 years. BIOCLIM Project, TA 5/10: Broad narrative descriptions of biosphere system changes in climate Transitions. CIEMAT/DIAE/551/55160/01/03.

Redmayne, D.W. and Musson, R.M.W. 1987. The Dunoon earthquake of 16 September 1985. Global Seismology Report No. 311. British Geological Survey.

Rendell, H., Worsley, P., Green, F. and Parks, D. 1991. Thermoluminescence dating of the Chelford Interstadial, Earth and Planetary Science Letters, 103, 182-189.

Richardson, S., Musson, R., & Horsburgh, K. 2007. Tsunami-assessing the hazard for the UK and Irish coast. Study commissioned by Defra Flood Management http://www.defra.gov.uk/environment/flooding/risk/tsunami.htm

Riggins, S.G., Anderson, R.S., Anderson, S.P. and Tye, A.M. 2011. Solving a Conundrum of a Steady-State Hillslope with Variable Soil Depths and Production Rates, Bodmin Moor, UK. Geomorphology. 128, 73-84.

Rignot E., I. Velicogna, M. R. van den Broeke, A. Monaghan, and J. Lenaerts (2011). Acceleration of the contribution of the Greenland and Antarctic ice sheets to sea level rise. *Geophysical Research Letters* **38** (5). L05503 Bibcode 2011GeoRL..3805503R. doi:10.1029/2011GL046583

Ringrose, P.S., Hancock, P.L., Fenton, C. and Davenport, C.A., 1991. Quaternary tectonic activity in Scotland. In: Foster, A., Culshaw, M.G., Cripps, J.C., Little, J.A., Moon, C.F., (Eds.) Quaternary Engineering Geology. Geological Society, Engineering Geology Special Publication, Vol. 7, pp. 679-686.

Ritchie, M.E.A., Musson, R.M.W. and Woodcock, N.H. 1990. The Bishop's Castle (UK) Earthquake of 2 April 1990. Terra Nova, 2/4, 390-400.

Richie, JD; Johnson, H and Kimbell, GS; 2003 The nature and age of Cenozoic contractional deformation within the NE Faroe-Shetland Basin ; Marine and Petroleum Geology; Vol. 20; No. 5; pp 399-409

Roberts, D. G. 1975. Marine Geology of the Rockall Plateau and Trough. Philosophical Transactions of the Royal Society of London, A278, 447-509.

Roberts, D. G. 1989. Basin inversion in and around the British Isles. In: Cooper, M. A. and Williams, G. D. (eds) Inversion Tectonics. Geological Society, London, Special Publication, 44, 131-150.

Roberts, D.G., Thompson, M., Mitchener, B., Hossack, J., Carmichael, S. and Bjørnseth, H.-M. 1999. Palaeozoic to Tertiary rift and basin dynamics: mid-Norway to the Bay of Biscay—a new context for hydrocarbon prospectivity in the deepwater frontier. In: Fleet, A.J. and Boldy, S.A.R. (Eds.), Petroleum Geology of Northwest Europe: Proceedings of the Fifth Conference. Geological Society, London, 7–40.

Rochelle, C.A. and Long, D. 2009. Gas hydrate stability in the vicinity of a deep geological repository for radioactive wastes: A scoping study. British Geological Survey Internal Report, OR/08/073. 24pp.

Rohrman, M. and Van der Beek, P. 1996. Cenozoic postrift domal uplift of North Atlantic margins: an asthenospheric diapirism model. Geology, 24, 901–904.

Rojstaczer, S., Wolf, S. and Michel, R., 1995. Permeability enhancement in the shallow crust as a cause of earthquake-induced hydrological changes, Nature 373, 237–239.

Rolfe, C.J. Hughes, P.D. Fenton, C.R. Schnabel, C. Xu, S. and Brown, A.G. 2012. Paired 26Al and 10Be exposure ages from Lundy: new evidence for the extent and timing of Devensian glaciation in the southern British Isles, Quaternary Science Reviews, 43, 61-73.

Rosa, E., Gaillardet, J., Hillaire-Marcel, C., Helie, J.F., Richard, L.F. 2012. Rock denudation rates and organic carbon exports along a latitudinal gradient in the Hudson, James and Ungava bays watershed. Canadian Journal of Earth Sciences, 49, 742-757.

Rose, J. 1985. The Dimlington Stadial/Dimlington Chronozone: a proposal for naming the main glacial episode of the Late Devensian in Britain. Boreas, 14: 225–230.

Rose, J. Moorlock, B. S. P. and Hamblin, R. J. O. 2001. Pre-Anglian fluvial and coastal deposits in Eastern England: lithostratigraphy and palaeoenvironments, Quaternary International, 79, 5-22.

Rose, J., Whiteman, C. A., Allen, P. and Kemp, R. A. 1999. The Kesgrave Sands and Gravels: 'pre-glacial' Quaternary deposits of the River Thames in East Anglia and the Thames valley, Proceedings of the Geologists' Association, 110, 2, 93-116.

Roussel, E.G., Cambon Bonavita, M-A., Querellou, J., Cragg, B.A., Webster, G., Prieur D., and Parkes, R.J., 2008, Extending the sub-sea-floor biosphere. Science, 20, p.1046.

Rowe, P. J., Atkinson, T. C. and Turner, C. 1999. U-series dating of Hoxnian interglacial deposits at Marks Tey, Essex, England. Journal of Quaternary Science, 14: 693–702

Ruddiman, W.F. 2003a. Orbital insolation, ice volume, and greenhouse gases .Quaternary Science Reviews, 22, 1597-1629.

Ruddiman, W. F. 2003b. The anthropogenic greenhouse era began thousands of years ago. Climatic Change, 61, 261-293.

Ruddiman, W. F. 2005. Ploughs, Plagues and Petroleum: How Humans Took Control of Climate. Princeton University Press.

Ruddiman, W. F., and Kutzbach, J. E. 1991. Plateau uplift and climatic change. Scientific American, 264, 66–75.

Ruddiman, W. F., Raymo, M. E., Martinson, D. G., Clement, B. M. and Backman, J. 1989. Pleistocene evolution: northern hemisphere ice sheets and North Atlantic Deep-Sea Core V23–82: correlation with the terrestrial record. Palaeoceanography, 4, 353–412.

Rudlof, W. 1981.World-climates, Wissenschaftliche Verlagsgesellschaft mbH.

Ruleman, C.A. and Lageson, D.R. 2002. Late Quaternary tectonic activity along the Madison fault zone, southwest Montana (abstract). Geological Society of America, Rocky Mountain 54th Annual Meeting, Cedar City, Utah (May 7–9, 2002).

Ruskeeniemi, T., Paananen, M., Ahonen, L., Kaija, J., Kuivamäki,A., Frape, S., Moren, L. and Degnan, P. 2002. Permafrost at Lupin: Report of Phase !. Geological Survey of Finland (GTK) Report. YST-112.

Samson, L. and Tordon, F. 1969. Experience with engineering site investigations on Northern Quebec and Northern Baffin Island. In Proceedings of the 3rd Candian Conference on Permafrost, January 14-15, pp 21-27.

Sargeant, S.L., Stafford, P.J., Lawley, R., Weatherill, G., Weston, A.-J.S., Bommer, J.J., Burton, P.W., Free, M., Musson, R.M.W., Kuuyuor, T. and Rossetto, T., 2008. Observations from the Folkestone, U.K., Earthquake of 28 April 2007, Seismol. Res. Lett., 79, 672-687

Sathar, S., Reeves, H.J., Cuss, R.J. and Harrington, J.F. 2012 The role of stress history on the flow of fluids through fractures. Mineralogical Magazine, 76, 8, 295-307.

Sauber, J. and Molnia, B.F. 2004. Glacier ice mass fluctuations and fault instability in tectonically active southern Alaska. Global and Planetary Change, 42, 279–293.

Sauber, J., Plafker, G., Molnia, B.F. and Bryant, M.A. 2000. Crustal deformation associated with glacial fluctuations in the eastern Chugach Mountains, Alaska. Journal of Geophysical Research, 105, 8055–8077.

Schatz, T., Martikainen, J., and Koskinen, K. 2010. Effect of freezing and thawing on compacted bentonite buffer performance. Clays in Natural and Engineered Barriers for Radioactive Waste Confinement, 4th International Meeting. March 2010, Nantes, France

Scherneck, H.-G., Johansson, J.M., Mitrovica, J.X. and Davis, J.L. 1998. The BIFROST project: GPS determined 3-D displacement rates in Fennoscandia from 800 days of continuous observations in the SWEPOS network. Tectonophysics, 294, 305–321.

Scherneck, H.-G., Johansson, J.M., Elgered, G., Davis, J.L., Jonsson, B., Hedling, G., Koivula, H., Ollikainen, M., Poutanen, M., Vermeer, M., Mitrovica, J.X. and Milne, G.A. 2000. BIFROST: Observing the Three-Dimensional Deformation of Fennoscandia. In: Mitrovica J.X. & Vermeersen, B.L.A. (eds) Glacial Isostatic Adjustment and the Earth System. Geodynamics Series, **29**, American Geophysical Union, Washington, D.C.

Scherneck, H.-G., Johansson, J.M., Vermeer, M., Davis, J.L., Milne, G.A. and Mitrovica, J.X. 2001. BIFROST Project: 3-D Crustal Deformation Rates Derived from GPS Confirm Postglacial Rebound in Fennoscandia. Earth Planets Space, 53, 703–708.

Schmidt, V. and McDonald, D.A. 1979. The role of secondary porosity in the course of sandstone diagenesis. In: Scholle, P.A. and Schluger, P.R. (editors). Aspects of Diagenesis Society of Economic Palaeontoogists and Mineralogists, Special Publication, 26, 175-207.

Scholz, C.H., 1998. Earthquakes and Friction Laws, Nature, 391, 37-42.

Schwartsev, S.L., Zuev, V.A. and Bukaty, M.B. 1988. Hydrochemistry of kryolithozone of Siberian Platform. In: proceedings of the 5th International Conference on permafrost, Trondheim, Norway, August2-5, 462-466.

Scourse, J. D., Haapaniemi A. I., Colmenero-Hidalgo, E., Peck, V. L., Hall, I. R., Austin, W. E. N., Knutz, P. C. and Zahn, R. 2009. Growth, dynamics and deglaciation of the last British–Irish ice sheet: the deep-sea ice-rafted detritus record, Quaternary Science Reviews, 28, 3066-3084.

Sébrier, M., Ghafiri, A. and Bles, J.L. 1997. Paleoseismicity in France: fault trench studies in a region of moderate seismicity. Journal of Geodynamics, 24, 207-217.

Seed, H. and Lee. K.L., 1966. Liquefaction of saturated sands during cyclic loading. Journal of the Soil Mechanics and Foundation Division 92: 105–134.

Seidov, D. and Haupt, B. J. 1997. Global ocean thermohaline conveyor at present and in the Late Quaternary. Geophysical Research Letters. 24, 2817-2820.

Sella, G.F., Stein, S., Dixon, T.H., Craymer, M., James, T.S., Mazzotti, S. and Dokka, R.K. 2007. Observations of glacial isostatic adjustment in the stable North America with GPS. Geophysical Research Letters, 34, L02346.

Sejrup, H. P., Hjelstuen, B. O., Dahlgren, K. I .T., Haflidason, H., Kuijpers, A., Nygård, A., Praeg, D., Stoker, M. S. and Vorren, T. O. 2005.Pleistocene glacial history of the NW European continental margin, Marine and Petroleum Geology, 22, 1111-1129.

Sejrup, H.P., Larsen, E., Landvik, J., King, E.L., Haflidason, H. and Nesje A. 2000. Quaternary glaciations in southern Fennoscandia: evidence from southwestern Norway and the northern North Sea region, Quaternary Science Reviews, 19, 667-685.

Serco, 2011 The Treatment of Geosphere Evolution in Post-Closure Performance Assessments. SERCO/TCS/002783/001 Issue 4,

Shackleton, N. J., Berger, A. and Peltier, W. R. 1990. An alternative astronomical calibration of the lower Pleistocene timescale based on ODP Site 677. Transactions of the Royal Society of Edinburgh: Earth Sciences, 81, 251-261.

Sharp, M., Tyranter, M., Brown, G.H. and Skidmore, M. 1995. Rates of chemical denudation and CO_2 drawdown in a glaciar-covered Alpine catchment. Geology, 23,61-64.

Shannon, P.M., Stoker, M.S., Praeg, D., van Weering, T.C.E., de Haas, H., Nielsen, T., Dahlgren, K.I.T., Hjelstuen, B.O., 2005. Sequence stratigraphic analysis in deep-water, underfilled NW European passive margin basins. Marine and Petroleum Geology, Vol 22; issues 9-10; doi: 10.1016/j.marpetgeo.2005.03.013.

Shennan, I. 1989, Holocene crustal movements and sea-level changes in Great Britain. J. Quaternary Sci., 4: 77–89.

Shennan, I., Hamilton, S., Hillier, C., Hunter, A., Woodall, R., Bradley, S., Milne, G., Brooks, A. and Basset, S. 2006a. Relative sea-level observations in western Scotland since the Last Glacial Maximum for testing models of glacial isostatic land movements and ice-sheet reconstructions, Journal Quaternary Science, 21, 601–613.

Shennan, I., Bradley, S., Milne, G., Brooks, A., Bassett, S. and Hamilton, S. 2006b. Relative sea-level changes, glacial isostatic adjustment modelling and ice-sheet reconstructions from the British Isles since the Last Glacial Maximum, J. Quat. Sci., 21, 585–599

Shennan, I. and Horton, B. 2002. Holocene land- and sea-level changes in Great Britain, Journal Quaternary Science, 17, 511–526.

Shennan, I., Lambeck, K., Horton, B., Innes, J., Lloyd, J., Mcarthur, J., Purcell, T. and Rutherford, M. 2000. Late Devensian and Holocene relative sea-level changes in northwest Scotland and their implications for glacio-hyrdro-isostatic modelling, Quaternary Science Reviews, 19, 1103–1135.

Shennan, I., Milne, G. and Bradley, S. 2009. Late Holocene relative land- and sea-level changes: Providing information for stakeholders. Geological Society of America Journal, 19(9), 52-53.

Shennan, I., Milne, G. and Bradley, S. 2012, Late Holocene vertical land motion and relative sea-level changes: lessons from the British Isles. Journal of Quaternary Science, 27: 64–70.

Shennan, I. and Horton, B. 2002. Holocene land- and sea-level changes in Great Britain. J. Quaternary Sci., 17: 511–526

Shennan, I., Peltier, W.R., Drummond, R. and Horton, B.P., 2002. Global to local scale parameters determining relative sea-level changes and the post-glacial isostatic adjustment of Great Britain, Quat. Sci. Rev., 21, 397–408

Shennan, I. and Woodworth, P.L. 1992. A comparison of Late Holocene and 20th Century sea-level trends from the UK and North Sea region. Geophysical Journal International, 109, 96-105.

Shepherd, A., Ivins, E. R., Geruo A., Barletta, V. R., Bentley, M. J., Bettadpur, S., Briggs, K. H., Bromwich, D. H., Forsberg, R., Galin, N., Horwath, M., Jacobs, S., Joughin, I., King, M. A., Lenaerts, J. T. M., Li, J., Ligtenberg, S. R. M., Luckman, A., Luthcke, S. B., McMillan, M., Meister, R.Milne, G., Mouginot, J., Muir, A. Nicolas, J. P., Paden, J., Payne, A. J., Pritchard, H., Rignot, E., Rott, H., Sandberg Sørensen, L., Scambos, T. A., Scheuchl, B., Schrama, E. J. O., Smith, B., Sundal, A. V., van Angelen, J. H., van de Berg, W. J.van den Broeke, M. R.,Vaughan, D. G., Velicogna, I., Wahr, J., Whitehouse, P. L., Wingham, D.J., Yi, D., Young, D.and. Zwally, H. J. 2012.A reconciled estimate of ice sheet mass balance. Science, 338, 1183-1189.

Sherwood Lollar, B. 2011. Far-field Microbiology Considerations Relevant to a Deep Geological Repository – State of Science Review. Nuclear Waste Management Organisation Review. NWMO TR-2011-09

Shur, Y. L., K. M. Hinkel, and F. E. Nelson 2005, The transient layer: Implications for geocryology and climate change science, Permafrost Periglacial Processes, 16, 5– 17.

Sissons, J.B. 1966. Relative sea-level changes between 10,300 and 8300 B.P in part of the Carse of Stirling. Transactions Institute British Geographers, 39, 19–29.

Sissons, J. B., Smith. D. E. and Cullingford, R. A. 1966. Lateglacial and postglacial shorelines in southeast Scotland. Transactions of the Institute of British Geographers, 39, 9-18.

SKB. 2006a. Long-term safety for KBS-3 repositories at Forsmark and Laxemar – a first evaluation: main report of the SR-Can project. TR-report SKBTR-06-09. Stockholm Sweden: Swedish Nuclear Fuel and Waste Management Company.

SKB 2006b. Climate and climate-related issues for the safety assessment SR-Can. SKB TR-06-23, Svensk Kärnbränslehantering AB.

SKB 2010. Climate and climate-related issues for the safety assessment SR-Site. SKB TR-10-49. Svensk Kärnbränslehantering AB. 328 pp.

SKB 2011 Long-term safety for the final repository for spent nuclear fuel at Forsmark. Main report of the SR-Site project. Updated 2011-12. Report number: KB TR-11-01.

Skinner, B. J. and Porter, S. C. 1995. The Dynamic Earth. (Third edition), Wiley: New York.

Small, E.E., Anderson, R.S. and Hancock, G.S., 1999. Estimates of the rate of regolith production using 10Be and 26Al from an alpine hillslope. Geomorphology, 27: 131-150.

Smistad, E.T., Tynan, M.C. and Swift, P.N., 2001. Consideration of disruptive events for the Yucca Mountain site recommendation report, Proc. WM01 Conf., 25 Feb-1 Mar 2001, Tucson AZ.

Solomon, S., Qin, D., Manning, M., Chen, Z,. Marquis, M.. Averyt, K.B., Tignor, M. and Miller, H. L (eds.). 2007. Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change. IPCC Fourth Assessment Report (AR4). (Cambridge University Press).

Spada, G., Yuen, D.A., Sabadini, R. and Boschi, E. 1991. Lower-mantle viscosity constrained by seismicity around deglaciated regions. Nature, 351, 53-55.

Stein, S., Sleep, N.H., Geller, R.J., Wang, S.C. and Kroeger G.C., 1979. Earthquakes along the passive margin of Eastern Canada, Geophys. Res. Lett., 6, 537–540.

Stein, S., Cloetingh, S., Sleep, N.H. and Wortel, R., 1989. Passive margin earthquakes, stresses and rhelogy. In: Gregersen, S., Basham, P.W. (Eds.), Earthquakes at North-Atlantic Passive Margins: Neotectonics and Postglacial Rebound. Series C: Mathematical and Physical Sciences. Kluwer Academic Publishers, 231–260.

Stepp, J. C., Wong, I., Whitney, J., Quittmeyer, R., Abrahamson, N., Toro, G., Youngs, R., Coppersmith, K., Savy, J. and Sullivan, T. (2001). Probabilistic seismic hazard analyses for ground motions and fault displacement at Yucca Mountain, Nevada. Earthquake Spectra, 17(1), 113-151.

Stevens, T. 1997. Lithoautotrophy in the subsurface. FEMS Microbiology Reviews. 20: 327-337.

Stewart, I.S., Sauber, J. and Rose, J., 2000. Glacio-seismotectonics: Ice sheets, crustal deformation and seismicity, Quat. Sci. Rev., 19, pp. 1367-1389.

Stoker, M.S., 2002. Late Neogene development of the UK Atlantic margin. In: Doré, A.G., Cartwright, J.A., Stoker, M.S., Turner, J.P. and Mite, N. (Eds.), Exhumation of the North Atlantic Margin: Timing, Mechanisms and Implications for Petroleum Exploration. Geological Society, London, Special Publications, 196, 411–438.

Stoker, M.S., Nielsen, T., van Weering, T.C.E. and Kuijpers, A. 2002. Towards an understanding of the Neogene tectonostratigraphic framework of the NE Atlantic margin between Ireland and the Faroe Islands. Marine Geology, 188, 233–248.

Stoker, M.S., and Shannon, P.M. 2005. Neogene evolution of the NW European Atlantic margin: Results from the STRATAGEM project (Editorial). Marine and Petroleum Geology, 22, 965-968.

Stoker, M.S., Praeg, D., Hjelstuen, B.O., Laberg, J.S., Nielsen, T., Shannon, P.M. 2005b. Neogene stratigraphy and the sedimentary and oceanographic development of the NW European Atlantic margin. Marine and Petroleum Geology, doi: 10.1016/j.marpetgeo.2004.11.007.

Stoker, M.S., Praeg, D., Shannon, P.M., Hjelstuen, B.O., Laberg, J.S., Nøttvedt, T., van Weering, T.C.E., Sejrup, H.P., Evans, D., 2005a. Neogene evolution of the Atlantic continental margin of NW Europe (Lofoten Islands to SW Ireland): anything but passive. In: Dore´, A.G., Vining, B. (Eds.), Petroleum Geology: NW Europe and Global Perspectives: Proceedings of the 6th conference. Geological Society, London, pp. 1057–1076.

Stoker, M.S., Praeg, D., Hjelstuen, B.O., Laberg, J.S., Nielsen, T., Shannon, P.M., 2005a. Neogene stratigraphy and the sedimentary and oceanographic development of the NW European Atlantic margin. Marine and Petroleum Geology, this issue, doi:10.1016/j.marpetgeo.2004.11.007.

Stoker, M.S., Hoult, R.J., Nielsen, T., Hjelstuen, B.O., Laberg, J.S., Shannon, P.M., Praeg D., Mathiesen, A., van Weering, T.C.E. McDonnell, A., 2005b. Sedimentary and oceanographic responses to early Neogene compression on the NW European margin. Marine and Petroleum Geology, this issue, doi:10.1016/j.marpetgeo.2005.01.009.

Stotler, R.L., Frape, S.K., Ruskeeniemi, T., Ahonen, L., Onstott, T.C. and Hobbs, M.Y. 2009. Hydrogeochemistry of groundwaters in and below the base of thick permafrost at Lupin, Nunavut, Canada. Journal of Hydrology, 373, 80-95.

Stotler, R.L., Frape, S.K., Ruskeeniemi, T., Ahonen, L., Onstott, T.C. and Hobbs, M.Y. 2009. Hydrogeochemistry of groundwaters in and below the base of thick permafrost at Lupin, Nunavut, Canada. Supplementary Information – doi:10.1016/j.jhydrol.2009.04.013.

Stroes-Gascoyne, S., Schippers A., Schwyn, B., Poulain, S., Sergeant, C., Simonoff, M., Le Marrec, C., Altmann, S., Nagaoka, T., Mauclaire, L., McKenzie, J., Daumas, S., Vinsot, A., Beaucaire, C. and Matray, J.-M. 2007. Microbial Community Analysis of Opalinus Clay Drill Core Samples from the Mont Terri Underground Research Laboratory, Switzerland. Geomicrobiology Journal. 24: 1-17.

Stroeven , A.P., D. Fabel, C. Hättestrand and J. Harbor 2002: A relict landscape in the centre of Fennoscandian glaciation: Cosmogenic radionuclide evidence of tors preserved through multiple glacial cycles. Geomorphology, 44 (1-2): 145-154.

Stouthamer, E. and, Berendsen, H.J.A. 2000. Factors controlling the Holocene avulsion history of the Rhine–Meuse delta (The Netherlands). Journal of Sedimentary Research, 70, 1051–1064.

Strong, G.E. and Milodowski, A.E. 1987. Aspects of the diagenesis of the Sherwood Sandstones of the Wessex Basin and their influence on reservoir characteristics. In: Marshall, J.D. (editor). Diagenesis of Sedimentary Sequences, Geological Society Special Publication No. 36, 325-337.

Sumbler, M. G., Bridgland, D. R., Wymer, J. J. and Balson, P. S. 1996. Quaternary, 110-136, in Sumbler M. G., British Regional Geology: London and the Thames Valley (4th edition) (London: HM SO for the British Geological Survey).

Summerfield, M.A., Stuart, F.M. Cockburn, H.A.P., Sugden, D.E., Denton, G.H., Dunai, T., Marchant, D.R. 1999. Long term denudation in the Dry Valleys, Transantarctic Mountains, southern Victoria Land, Antarctica based on in-situ-produced cosmogenic Ne-21. Geomorphology, 27, 113-129.

Sutherland, D. G. 1993. Muckle Head and Selwick. 74-76, in Quaternary of Scotland. Gordon, J E, and Sutherland, D. G. (editors) Geological Conservation Review Series, 6. (London: Chapman and Hall).

Sundquist, E.T., 1990: 'Long-term aspects of future atmospheric CO2 and sea level changes', In Revelle, R. (ed.), Sea-level Change, National Research Council Studies in Geophysics, National Academy Press, Washington D.C., pp.193-207.Sutherland, D. G. and Gordon, J. E. 1993. The Quaternary in Scotland. 13-47, in Quaternary of Scotland. Gordon, J E, and Sutherland, D. G. (editors) Geological Conservation Review Series, 6. (London: Chapman and Hall).

Swiecicki, T., Gibbs, P.B., Farrow, G.E. and Coward, M.P. 1998. A tectonostratigraphic framework for the mid-Norway region. Marine and Petroleum Geology, 15, 245–276.

Talbot, C.J. 1999. Ice ages and nuclear waste isolation. Engineering Geology, 52, 177-192.

Talbot, C.J. and Slunga, R. 1989. Patterns of Active Shear in Fennoscandia. In: GREGERSEN, S. and BASHAM, P.W. (Eds.), Earthquakes at North Atlantic passive Margins: Neotectonics and Postglacial Rebound. Kluwer Academic Publisher, Dordrecht, 441-466.

Tamisiea, M.E., Mitrovica, J.X. and Davis, J.L. 2007. GRACE gravity data constrain ancient ice geometries and continental dynamics over Laurentide, Science, 316, 881–883.

Tate, M.P. 1993. Structural framework and tectonostratigraphic evolution of the Porcupine Seabight Basin, offshore western Ireland. Marine and Petroleum Geology, 10, 95–123.

Tate, M.P., Dodd, C.D. and Grant, N.T. 1999. The Northeast Rockall Basin and its significance in the evolution of the Rockall-Faroes/East Greenland rift system. In: Fleet, A.J. and Boldy, S.A.R. (Eds.), Petroleum Geology of Northwest Europe: Proceedings of the Fifth Conference. Geological Society, London, 391–406.

Teferle, F. N., Bingley, R. M., Orliac, E. J., Williams, S. D. P., Woodworth, P. L., McLaughlin, D., Baker, T. F., Shennan I., Milne, G. A., Bradley, S. L. and Hansen D. N. 2009. Crustal motions in Great Britain: evidence from continuous GPS, absolute gravity and Holocene sea level data. Geophysical Journal International.

Texier, D., Degnan, P,, Loutre, M. F., Paillard, D. and Thorne, M. 2003. Modelling Sequential BIOsphere Systems under CLIMate Change.for Radioactive Waste Disposal. Project

Thiede, .J, Jessen, C., Knutz, P., Kuijpers, A Mikkelsen, N., Norgaard-Pedersen, N. and Spielhagen, R. 2011. Millions of Years of Greenland Ice Sheet History Recorded in Ocean Sediments. Polarforschung. 80(3), 141-159.

Thomas, H.R., Cleall, P., Li, Y.-C., Harris, C. and Kern-Luetschg, M. 2009 Modelling of cryogenic processes in permafrost and seasonally frozen soils. Géotechnique, 59 (3), 173-184, doi: 10.1680/geot.2009.59.3.173.

Thorne, M.C. and Kane, P. 2006. Development of a series of narratives for climatic and landscape change, Mike Thorne and Associates Limited Report to United Kingdom Nirex Limited MTA/P0011a/2005-1: Issue 2.

Thorne, M.C. 2010. NDA RWMD Biosphere Assessment Studies FY2009-2010: Representation of climate states and climate state transitions taking key FEPs into account, Quintessa Limited report to the Nuclear Decommissioning Authority QRS-1378K-3, Version 2.0

Torbjørn Dahlgren, K.I., Vorren, T.O., Stoker, M.S., Nielsen, T., Nygård, A. and Sejrup, H.P. 2005. Late Cenozoic prograding wedges on the NW European continental margin: their formation and relationship to tectonics and climate. Marine and Petroleum Geology, 22, 1089-1110.

Tiley, R, White, N, and Al-Kindi, S; 2004 Linking Paleogene denudation and magmatic underplating beneath the British Isles Geol. Mag. 141 (3), 2004, pp. 345–351 DOI: 10.1017/S0016756804009197.

Toynton, R. 1983**.** The relation between fracture patterns and hydraulic anisotropy in the Norfolk Chalk, England. Quarterly Journal of Engineering Geology*,* 16, 169-85.

Tuck, V. M., Edyvean, R. G. J., West, J. M., Bateman, K., Coombs, P., Milodowski, A. E. and McKervey, J. A. 2006 Biologically induced clay formation in subsurface granitic environments. J Geochem Explor 90 (1-2), 123-133. doi.org/10.1016/j.gexplo.2005.09.007.

Turbitt, T., Barker, E.J., Browitt, C.W.A., Howells, M., Marrow, P.C., Musson, R.M.W., Newmark, R.H., Redmayne, D.W., Walker, A.B., Jacob, A.W.B., Ryan, E. and Ward, V., 1985. The North-Wales Earthquake of 19 July 1984. J. Geol. Soc. (London) 142, 567–571.

Tullborg, E-L., Landström, O. and Wallin, B., 1999. Low-temperature trace element mobility influenced by microbial activity - indications from fracture calcite and pyrite in crystalline basement. Chemical Geology, 157: 199-218.

Turner, C. 1970. The Middle Pleistocene deposits at Marks Tey, Essex. Philosophical Transactions of the Royal Society, London, B257, 373–440.

Turpeinen, H., Hampel, A., Karow, T. and Maniatis, G. 2008. Effect of ice sheet growth and melting on the slip evolution of thrust faults. Earth and Planetary Science Letters, 269, 230–241.

Tushingham, A.M. and Peltier, W.R., 1991. ICE-3G—A new global-model of late Pleistocene deglaciation based upon geophysical predictions of postglacial relative sea-level change, J. Geophys. Res., 96, 4497–4523.

Tye, A.M., Kemp, S.J., Lark, R.M. and Milodowski, A.E. 2012. The role of peri-glacial active layer development in determining soil-regolith thickness across a Triassic sandstone outcrop in the UK. Earth Surface Processes and Landforms, 37(9), 971-983.

Tzedakis, P. C., Channell, J. E. T., Hodell, D. A., Kleiven, H. F. and Skinner, L. C. 2012. Determining the natural length of the current interglacial. Nature Geoscience, 5, 138–141.

UKCIP02, 2002. Climate Change Scenarios for the United Kingdom: The UKCIP02 Briefing Report, April 2002

Ustaszewski, M., Hampel, A. and Pfiffner, O.A. 2008. Composite faults in the Swiss Alps formed by the interplay of tectonics, gravitation and postglacial rebound: an integrated field and modelling study. Swiss Journal of Geosciences, 101, 223–235.

Vaikmae, E, R., Edmunds, W.M. and Manzano, M., 2001. Weichselian palaeoclimate and palaeoenvironment in Europe: background for palaeogroundwater formation. *From:* Edmunds, W. M. and Milne, C. J. (eds). 2001. *Palaeowaters in Coastal Europe: evolution of groundwater since the late Pleistocene.* Geological Society, London, Special Publications, 189, 163-191.

Vågnes, E., Gabrielsen, R.H. and Haremo, P. 1998. Late Cretaceous—Cenozoic intraplate contractional deformation at the Norwegian continental shelf: timing, magnitude and regional implications. Tectonophysics, 300, 29–46.

Van Balen, R.T., Houtgast, R.F., Van der Wateren, F.M., Vandenberghe, J. and Bogaart, P.W. 2000. Sediment budget and tectonic evolution of the Meuse catchment in the Ardennes and the Roer Valley Rift System. Global and Planetary Change, 27, 113–129.

Vanneste, K; Henriet, J-P; Posewang, J and Theilen, F; 1995; Seismic stratigraphy of the Bill Bailey and Lousy Bank area: Implications for subsidence history. In: Scrutton, RA; Stoker, MS; Shimmield, GB and Tudhope, AW (eds) The Tectonics, Sedimentation and Palaeoceanography of the North Atlantic Region; Geological Society of London Special Publication 90 p p125-139.

Van Geet, M., De Craen, M., Mallants, D., Wemaere, I., Wouters, L. and Cool, W. 2009. How to treat climate evolution in the assessment of the long-term safety of disposal facilities for radioactive waste: examples from Belgium. Climate of the Past Discussions 5, 463–494.

Van Vliet-Lanoë , B., Laurent, M., Everaerts, M., Mansy, J.-L. and Manby, G. 2000. Evolution Neogene et quarternaire de la Somme, une flexuration tectonique active. Earth and Planetary Sciences, 331, 151–158.

Van Vliet-Lanoë, B., Vandenberghe, N., Moguedet, G. and Michel, Y., 2002. Palaeogeographic evolution of northwestern Europe during the Upper Cenozoic. Geodiversitas, 24 (3), 511–541.Vanneste, K., Henriet, J.-P., Posewang, J. and Theilen, F. 1995. Seismic stratigraphy of the Bill Bailey and Lousy Bank area: implications for subsidence history. In: Scrutton, R.A., Stoker, M.S., Shimmield, G.B. and Tudhope, A.W. (Eds.), The Tectonics, Sedimentation and Palaeoceanography of the North Atlantic Region, Geological Society, London, Special Publication, 90, 125–139.

van Weert, F. H. A., van Gijssel, K., Leijnse, A. and Boulton, G.S. 1997. The effects of Pleistocene glaciations on the geohydrological system of Northwest Europe. Journal of Hydrology, 195(1– 4), 137–159.

Varley, P.M. 1996. Seismic risk assessment and analysis, in Harris, Colin S., Engineering geology of the Channel Tunnel, Thomas Telford, pp. 195–8.

Vaughan-Hirsch, D.P., Phillips, E., Lee, J.R. and Hart, J.K. 2012. Micromrphological analysis of poly-phase deformation associated with the transport and emplacement of glaciotectonic rafts at West Runton, north Norfolk, UK. Boreas, DOI 10.1111/j.1502-3885.2012.00268.x

Velbel, M.A. 1985. Geochemical mass balances and weathering rates in forested watersheds of the southern Blue Ridge. American Journal of Science, 285, 904-930.

Velbel, M.A. 1986. The mathematical basis for determining rates of geochemical and geomorphic processes in small forested watersheds by mass-balance. Examples and implications. In Rates of Chemical Weathering of Rocks and Minerals Eds. Colman, S.M.; Dethier, D.P., 439-451.

Vernon, A.J., van der Beek, P.A., Sinclair, H.D., Rahn, M.K. 2008. Increase in late Neogene denudation of the European Alps confirmed by analysis of a fission-track thermochronology database. Earth and Planetary Science Letters, 270, 316-329.

Vidstrand P, Follin S, and Zugec N, 2010. Groundwater flow modelling of periods with periglacial and glacial conditions – SR-Site Forsmark. SKB R-09-21, Svensk Kärnbränslehantering AB.

Viville, D., Vhabaux, F., Stille, P., Pierret, M.C. and Gangloff, S. 2012. Erosion and weathering fluxes in granitic basins: The example of the Strengbach catchment (Vosges massif, eastern France). Catena, 92, 122-129.

Wahr, AGJ and Zhong, S; 2013 Computations of the viscoelastic response of a 3-D compressible Earth to surface loading: an application to Glacial Isostatic Adjustment in Antarctica and Canada; Geophys. J. Int., 192, 557–572, doi: 10.1093/gji/ggs030.

Walcott, R.I., 1970. Isostatic response to loading of the crust in Canada. Canadian Journal of Earth Sciences, 7, 716-729.

Walker, A.B. (ed). 1993. Bulletin of British earthquakes 1992. British Geological Survey Technical report WL/93/11.

Walker, A.B. (ed). 1994. Bulletin of British earthquakes 1993. British Geological Survey Technical report WL/94/9.

Walker, A.B. (ed). 1995. Bulletin of British earthquakes 1994. British Geological Survey Technical report WL/95/4.

Walker, M. J. C., Merritt, J. W., Auton, C. A., Coope, G. R., Field, M. H., Heijnis, H. and Taylor, B. J. 1992. Allt Odhar and Dalcharn: Two pre late-Devensian (late Weichselian) sites in Northern Scotland. Journal of Quaternary Science. 7, 69-86.

Walden, J., Wadsworth, E., Austin, W. E. N., Peters, C., Scourse, J. D. and Hall, I. R. 2006. Compositional variability of ice-rafted debris in Heinrich layers 1 and 2 on the northwest European continental slope identified by environmental magnetic analyses. J. Quaternary Sci., 22 163–172.

Walling, D.E. and Webb, B.W. 1987. Suspended load in gravel bed rivers: UK Experience. In Thorne, C.R., Bathurst, J.C. and Hay, R.D. (eds). Sediment Transport in Gravel Bed rivers.Joun Wiley and Sons, Chichester, pp. 691-723.

Wang, C.-Y., Cheng, L.-H., Chin, C.-V. and Yu S.-B., 2001. Coseismic hydrologic response of an alluvial fan to the 1999 Chi-Chi earthquake, Taiwan, Geology 29, 831–834.

Wang, C.Y., 2007. Liquefaction beyond the Near Field, Seismol. Res. Lett., 78, 5, 512-517

Wanner, H., Beer, J., Bütikofer, J., Crowley, T. J., Cubasch, U., Flückiger, J., Goosse, H., Grosjean, M., Joos, F., Kaplan, J. O., Küttel, M., Müller S. A., Prentice, I. C., Solomina, O., Stocker, T. F., Tarasov, P., Wagner, M. and Widmann, M. 2008. Mid- to Late Holocene climate change: an overview, Quaternary Science Reviews, 27, 1791-1828.

Wells, D.L. and Coppersmith, K.J., 1994. New empirical relationships among magnitude, rupture length, rupture width, rupture area, and surface displacement, Bull. Seismol. Soc. Am., 84, 974-1002.

Wersin, P., Stroes-Gascoyne, S., Pearson, F. J., Tournassat, C., Leupin, O. X. and Schwyn, B. 2011. Biogeochemical processes in a clay formation in situ experiment: Part G - Key interpretations and conclusions. Implications for repository safety. Applied Geochemistry. 26: 1023-1034

West JM, and Chilton PJ. 1997Aquifers as environments for microbiological activity. Quarterly Journal of Engineering Geology 30:147-154.

West, J M and McKinley I G 2002 The geomicrobiology of radioactive waste disposal. In: The Encyclopedia of Environmental Microbiology (Editor G. Bitton), John Wiley. pp 2661-2674 http://eu.wiley.com/WileyCDA/WileyTitle/productCd-0471354503.html.

West, J. M., McKinley, I. G. and Stroes-Gascoyne, S. 2002 Microbial effects on waste repository materials. In: Microbiology and Radioactivity (Editors M. Keith-Loach and F. Livens), Elsevier. pp 255-278. doi.org/10.1016/S1569-4860(02)80038-9.

West, R. O., Funnell, B. M., & Norton, P. E. (1980). An Early Pleistocene cold marine episode in the North Sea: pollen and faunal assemblages at Covehithe, Suffolk, England. Boreas, 9(1), 1-10.

Whitehouse, P., 2006. Chapter 3.3 Isostatic adjustment and shoreline migration. In: Climate and climate-related issues for the Safety Assessment SR-Can. TR-report SKB TR-06-23. Stockholm, Sweden: Swedish Nuclear Fuel and Waste Management Company. 68-92

Whitman W B, Coleman D C, Wiebe W J. 2001. Prokaryotes: The unseen majority. Proc. National Academy of Science, USA 95:6578-6583.

Wieczerkowski, K., Mitrovica, J.X. and Wolf, D. 1999. A revised relaxation time spectrum for Fennoscandia. Geophysical Journal International, 139, 69-86.

Wildenborg, AFB, Orlic, B, Thimus, JF., de Lange, G, de Cock, S, de Leeuw, CS, and Veling, EJM, 2003. Radionuclide transport in clay during climate change, Netherlands Journal of Geosciences / Geologie en Mijnbouw, 82 (1): 19-30

Williams, G.A., Turner, J.P. and Holford, S.P. 2005. Inversion and exhumation of the St. George's Channel Basin, offshore Wales, UK. Journal of the Geological Society, London, 162, 97–110.

Wilkinson, M.T., 2005. Soil production in heath and forest, Blue Mountains, Australia: influence of lithology and palaeoclimate. Earth Surface Processes and Landforms, 30: 923-934.

Woodcock, N.H. and Strachan, R.A., 2000. Geological History of Britain and Ireland. John Wiley and Sons.

Woodland, A W 1970. The buried tunnel-valleys of East Anglia. Proceedings of the Yorkshire Geological Society 37: 521-578

Wragg, J., Harrison, H., West, J. M., Yoshikawa, H. 2012. Comparison of microbiological influences on the transport properties of intact mudstone and sandstone and its relevance to the geological disposal of radioactive waste. Mineralogical Magazine. 76(8), 3251-3259 doi: 10.1180/minmag.2012.076.8.38.

Wu, P., 1996. Changes in orientation of near-surface stress field as constraints to mantle viscosity and horizontal stress differences in Eastern Canada. Geophysical Research Letters, 23, 2263-2266.

Wu, P., 1997. Effects of viscosity structure on fault potential and stress orientations in eastern Canada. Geophysical Journal International, 130, 362-365.

Wu, P., 1999. Intraplate earthquakes and postglacial rebound in eastern Canada and Northern Europe. In: WU, P. (Ed.), Dynamics of the Ice Age Earth, a Modern Perspective. Transactions Technical Publication, Zurich, 443-458.

Wu, P. and Hasegawa, H. 1996a. Induced stresses and fault potential in eastern Canada due to a disc load - a preliminary analysis. Geophysical Journal International, 125, 415-430.

Wu, P. and Hasegawa, H. 1996b. Induced stresses and fault potential in eastern Canada due to a realistic load - a preliminary analysis. Geophysical Journal International , 125, 215-219.

Wu, P. and Johnston, P., 2000. Can deglaciation trigger earthquakes in N. America? Geophys. Res. Lett., 27, 1,323–1,326.

Wu, P., Johnston, P. and Lambeck, K. 1999. Postglacial rebound and fault instability in Fennoscandia. Geophysical. Journal International, 139. 657-670.

Wu, P. and Peltier, W.R. 1984. Pliestocene deglaciation and the Earth's rotation: A new analysis. Geophysical Journal of the Royal Astronomical Society, 76, 753-792.

Wymer, J, 1988. Palaeolithic archaeology and the British Quaternary sequence, Quaternary Science Reviews, 7, 79-98.

Youd, T.L., 2003. Liquefaction mechanisms and induced ground failure. In: Lee, W.H.K., Kanamori, H., Jennings, P.C. and Kisslinger, C. (eds.) International Handbook of Earthquake and Engineering Seismology, Part B, 1159–1173, Academic Press.

Zhukovskiy, S.Y., Mazurov, O.S. and Pirogov, I.A. 1973. Properties of the formation and quantitative estimate of the ice content of grantite permafrost at the construction site of the Kolyma hydroelectric power plant. In: Proceedings of the 2nd International Conference on Permafrost, Yakutsk, USSR, July 13-28, 1873, 665-668.

Ziegler, P. A. 1982. Geological Atlas of Western and Central Europe. Elsevier Science Publishers, Amsterdam.

Original sources of images used in this report

Figure 1. Modified after Evans, CJ; 1987; Crustal stress in the United Kingdom: British Geological Survey, Investigation of the geothermal potential of the U.K. report WJ/GE/87/008; 31p.

Figure 2. Modified after Assumpcao, M. 1981. The NW Scotland earthquake swarm of 1974. Geophysical Journal of the Royal Astronomical Society, 67, 577-586; Marrow, P.C. and Roberts, G. 1984. Source mechanisms of aftershocks of the 1979 'Carlisle' earthquake. Global Seismology Unit Report No. 217, British Geological Survey; Turbitt, T., Barker, E.J., Browitt, C.W.A., Howells, M., Marrow, P.C., Musson, R.M.W., Newmark, R.H., Redmayne, D.W., Walker, A.B., Jacob, A.W.B., Ryan, E. and Ward, V., 1985. The North-Wales Earthquake of 19 July 1984. J. Geol. Soc. (London) 142, 567–571; Redmayne, D.W. and Musson, R.M.W. 1987. The Dunoon earthquake of 16 September 1985. Global Seismology Report No. 311. British Geological Survey; Ritchie, M.E.A., Musson, R.M.W. and Woodcock, N.H. 1990. The Bishop's Castle (UK) Earthquake of 2 April 1990. Terra Nova, 2/4, 390-400; Walker, A.B. (ed). 1993. Bulletin of British earthquakes 1992. British Geological Survey Technical report WL/93/11; Walker, A.B. (ed). 1994. Bulletin of British earthquakes 1993. British Geological Survey Technical report WL/94/9 and Walker, A.B. (ed). 1995. Bulletin of British earthquakes 1994. British Geological Survey Technical report WL/95/4.

Figure 3. Reprinted from Cloetingh, S., Ziegler, P.A., Beekman, F., Andriessen, P.A.M., Matenco, L., Bada, G., Garcia-Castellanos, D., Hardebol, N., Dèzes, P. and Sokoutis, D. 2005. Lithospheric memory, state of stress and rheology: neotectonic controls on Europe's intraplate continental topography. Quaternary Science Reviews, 24, 241-304 with permission from Elsevier.

Figure 4. Reprinted from Cloetingh, S., Ziegler, P.A., Beekman, F., Andriessen, P.A.M., Matenco, L., Bada, G., Garcia-Castellanos, D., Hardebol, N., Dèzes, P. and Sokoutis, D. 2005. Lithospheric memory, state of stress and rheology: neotectonic controls on Europe's intraplate continental topography. Quaternary Science Reviews, 24, 241-304 with permission from Elsevier.

Figure 5 Orginal Figure

Figure 6. Image from "http://grace.jpl.nasa.gov. Wahr, AGJ and Zhong, S; 2013 Computations of the viscoelastic response of a 3-D compressible Earth to surface loading: an application to Glacial Isostatic Adjustment in Antarctica and Canada; Geophys. J. Int., 192, 557–572, doi: 10.1093/gji/ggs030

Figure 7. Reprinted from Stewart, I.S., Sauber, J. and Rose, J., 2000. Glacio-seismotectonics: Ice sheets, crustal deformation and seismicity, Quat. Sci. Rev., 19, pp. 1367-1389 with permission from Elsevier.

Figure 8. Reprinted from Stewart, I.S., Sauber, J. and Rose, J., 2000. Glacio-seismotectonics: Ice sheets, crustal deformation and seismicity, Quat. Sci. Rev., 19, pp. 1367-1389 with permission from Elsevier.

Figure 9. Reprinted from Stewart, I.S., Sauber, J. and Rose, J., 2000. Glacio-seismotectonics: Ice sheets, crustal deformation and seismicity, Quat. Sci. Rev., 19, pp. 1367-1389 with permission from Elsevier.

Figure 10. Lund, B. and Näsland, J.O. 2009. 5. Glacial isostatic adjsment: implications for glacially induced faulting and nuclear waste repositories. In: Connor, C.B., Chapman, N.A. and Connor, L.J. (eds), Volcanic and tectonic hazard assessment for nuclear facilities. Cambridge University Press, 142-155.

Figure 11. Lund, B. and Näsland, J.O. 2009. 5. Glacial isostatic adjsment: implications for glacially induced faulting and nuclear waste repositories. In: Connor, C.B., Chapman, N.A. and Connor, L.J. (eds), Volcanic and tectonic hazard assessment for nuclear facilities. Cambridge University Press, 142-155.
Figure 12: Musson, R.M.W., 1996. The seismicity of the British Isles. Annali di Geofisica 39, 463–469.

Figure 13: Baptie, B., 2010. Seismogenesis and state of stress in the UK. Tectonophysics, 482 (1-4), 150-159.

Figure 14: Baptie, B., 2010. Seismogenesis and state of stress in the UK. Tectonophysics, 482 (1-4), 150-159.

Figure 15: Musson, R.M.W. and Sargeant, S.L., 2007. Eurocode 8 seismic hazard zoning maps for the UK, British Geological Survey Technical Report CR/07/125

Figure 16: Original figure by Baptie, B.

Figure 17: Modified after Manga, M. and C.-Y. Wang, 2007. Earthquake hydrology, in Treatise on Geophysics, G. Schubert editor, 4, 293-320. Elsevier

Figure 18. Redrawn after Skinner, B. J. and Porter, S. C. 1995. The Dynamic Earth. (Third edition), Wiley: New York.

Figure 19 Redrawn after Raymo, M E. 1997. The timing of major climate terminations. Palaeooceanography, 12, 577-585.

Figure 20. Modified after Merritt, J. W., Auton, C. A., Connell, E. R, Hall, A. M. and Peacock, J. D. 2003. Cainozoic geology and landscape evolution of north-east Scotland. Memoir of the British Geological Survey, Sheets 66E, 67, 76E, 77, 86E, 87W, 87E, 95, 96W, 96E and 97 (Scotland).

Figure 21. Modified after NDA 2010 Geological Disposal: Geosphere status report; Nuclear Decommissioning Authority Report NDA/RWMD/035, 2010., prinicipally with data from reconstructions by Clark, J., McCabe, A. M., Bowen, D. Q. and Clark, P. U. 2012a. Response of the Irish Ice Sheet to abrupt climate change during the last deglaciation, Quaternary Science Reviews, 35, 100-115 and Gibbard, P. L. and Clark, C. D. 2011. Pleistocene glaciation limits in Great Britain. 75-93. In:. Elhers, J., Gibbard, P. L. and Hughes, P.D. (eds), Developments in Quaternary Science, 15: Quaternary Glaciations - Extent and Chronology: A Closer Look. Elsevier.

Figure 22. NASA image by Jeff Schmaltz, MODIS Rapid Response Team, Goddard Space Flight Center. Caption by Michon Scott.
http://eoimages.gsfc.nasa.gov/images/imagerecords/42000/42237/gbritain_tmo_2010007_lrg.jpg

Figure 23. Modified after Clark, C. D., Hughes, A. L.C., Greenwood, S. L., Jordan, C. and Sejrup, H. P. 2012b. Pattern and timing of retreat of the last British-Irish Ice Sheet, Quaternary Science Reviews, 44, 112-146.

Figure 24. Modified after Loutre, M F. and Berger, A. 2000. Future climatic changes: Are we entering an exceptionally long interglacial? Climatic Change, 46, 61-90.

Figure 25. Modified after Cedercreutz, J. 2004. Future climate scenarios for Olkiloto with emphais on permafrost. Posiva Report 2004-2006.

Figure 26. Modified after BIOCLIM 2001. Deliverable D3: Global climate features over the next million years and recommendation for specific situations to be considered.

Figure 27. Modified after from BIOCLIM 2003. Deliverable D7: Continuous climate evolution scenarios over Western Europe (1000 km scale).

Figure 28. Modified after BIOCLIM 2003c. Deliverable 8a: Development of the rule-based downscaling methodlogy for BIOCLIM Workpackage 3.

Figure 29.Modified after BIOCLIM 2003. Deliverable D7: Continuous climate evolution scenarios over Western Europe (1000 km scale).

Figure 30. Modified after Merritt, J. W., Auton, C. A., Connell, E. R, Hall, A. M. and Peacock, J. D. 2003. Cainozoic geology and landscape evolution of north-east Scotland. Memoir of the British Geological Survey, Sheets 66E, 67, 76E, 77, 86E, 87W, 87E, 95, 96W, 96E and 97 (Scotland).

Figure 31. Reprinted from Hubbard, A, Bradwell, T, Golledge, N, Hall, A, Patton, H, Sugden, D, Cooper, R and Stoker, M. 2009. Dynamic cycles, ice streams and their impact on the extent, chronology and deglaciation of the British–Irish ice sheet, Quaternary Science Reviews, 28, 758-776 with permission from Elsevier. 93

Figure 32. Modified after NDA 2010 Geological Disposal: Biosphere status report. Nuclear Decommissioning Authority Report NDA/RWMD/036, 2010.

Figure 33Modified after Bradley, S. L., Milne, G. A., Shennan, I. and Edwards, R. 2011. An improved glacial isostatic adjustment model for the British Isles. J. Quaternary Sci., 26: 541–552

Figure 34 Reprinted from figure 2 in Bradley, S. L., Milne, G. A., Teferle, F. N., Bingley, R. M. and Orliac, E. J. 2009. Glacial isostatic adjustment of the British Isles: new constraints from GPS measurements of crustal motion. Geophysical Journal International, 178, 14–22.

Figure 35 Modified after Goodess, C.M., Watkins, S.J., Palutikof, J.P. and Thorne, M.C. 2004. The construction of global eustatic sea-level scenarios for the next 150,000 years. Climatic Research Unit Research Paper Number 3 (Second Series) Version 4, October 2004

Figure 36 Modified after Goodess, C.M., Watkins, S.J., Palutikof, J.P. and Thorne, M.C. 2004. The construction of global eustatic sea-level scenarios for the next 150,000 years. Climatic Research Unit Research Paper Number 3 (Second Series) Version 4, October 2004

Figure 37. Modified after Hagdorn, M. K. M. 2003. Reconstruction of the Past and Forecast of the Future European and British Ice Sheets and Associated Sea–Level Change. (Unpublished PhD. Thesis, University of Edinburgh).

Figure 38 Figure taken from Millin, S. (editor). 2010. UKCP09 sea level change estimates. UKCIP Briefing notes, 7pp.

Figure 39 Modified after Evans, E., Ashley, R., Hall, J., Pennington-Roswell, E. Saul, A., Sayers, P., Thorne, C. and Watkinson, A. 2004. Foresight Future Flooding. Scientific Summary: Volume 1 – Future Risks and Their Drivers. (Office of Science and Technology, London).

Figure 40 Figure from PADAMOT project workshop

Figure 41 . Milodowski, A.E. Tullborg, E.-L., Buil, B., Gómez, P., Turrero, M.-J., Haszeldine, S., England, G., Gillespie, M.R., Torres, T., Ortiz, J.E., Zachariáš, Silar, J., Chvátal, M., Strnad, L., Šebek, O. Bouch, J.E., Chenery, S.R., Chenery, C., Shepherd, T.J. and Mckervey, J.A. 2005. Application of Mineralogical, Petrological and Geochemical tools for Evaluating the Palaeohydrogeological Evolution of the PADAMOT Study Sites. PADAMOT PROJECT - EU FP5 CONTRACT NO. FIKW-CT2001-20129. PADAMOT Project Technical Report, WP2. 228pp + digital appendices.

Figure 42. Milodowski, A.E. Tullborg, E.-L., Buil, B., Gómez, P., Turrero, M.-J., Haszeldine, S., England, G., Gillespie, M.R., Torres, T., Ortiz, J.E., Zachariáš, Silar, J., Chvátal, M., Strnad, L., Šebek, O. Bouch, J.E., Chenery, S.R., Chenery, C., Shepherd, T.J. and Mckervey, J.A. 2005. Application of Mineralogical, Petrological and Geochemical tools for Evaluating the Palaeohydrogeological Evolution of the PADAMOT Study Sites. PADAMOT PROJECT - EU FP5 CONTRACT NO. FIKW-CT2001-20129. PADAMOT Project Technical Report, WP2. 228pp + digital appendices.

Figure 43. Irish Sea processed bathymetry (copyright BGS/NERC
http://jncc.defra.gov.uk/pdf/irish_finalreport(maps).pdf)