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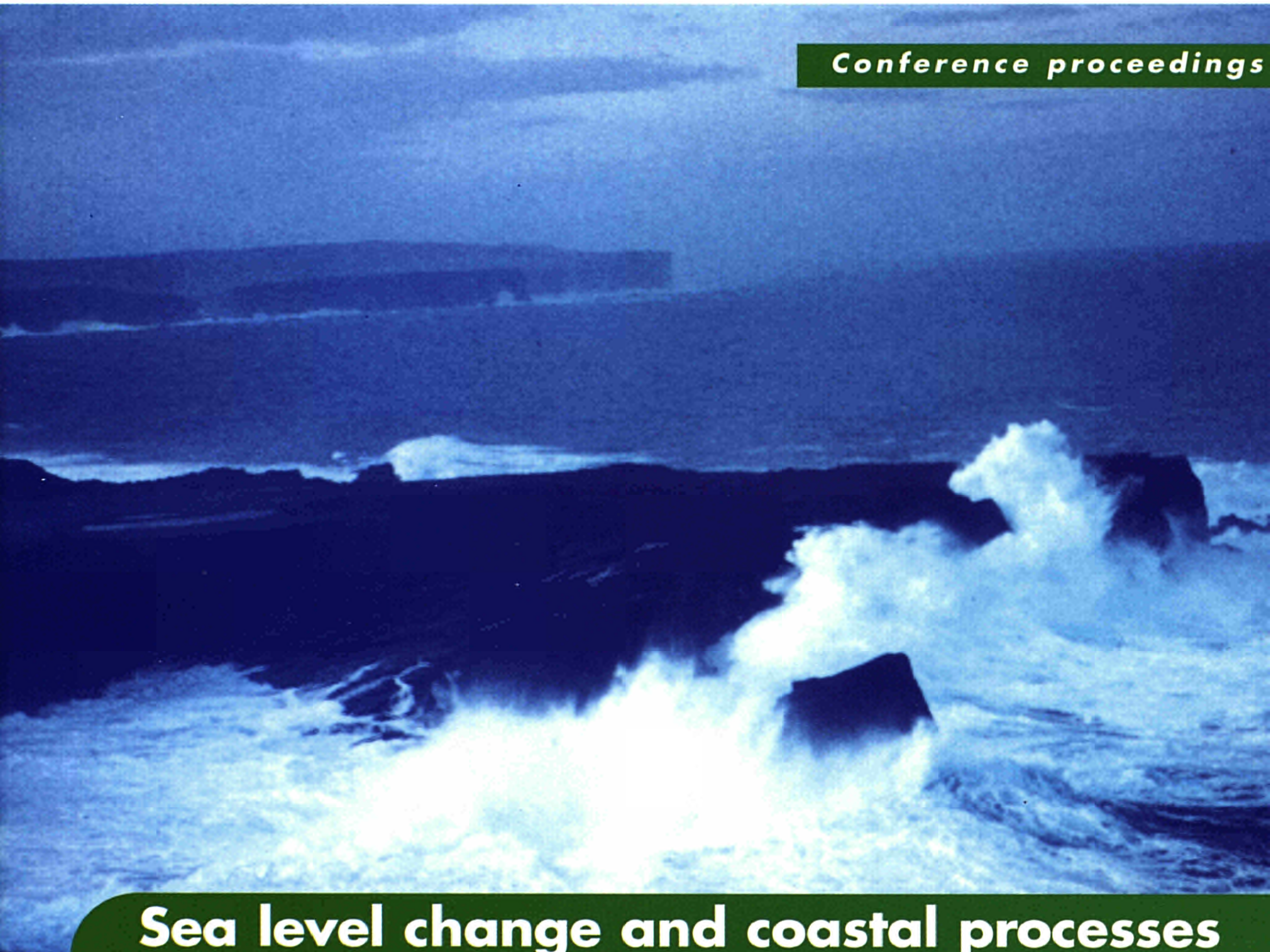

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Sea level change and coastal processes Implications for Europe

Research results and recommendations

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European Commission

Sea level change and coastal processes

Implications for Europe

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2000 EUR 19337

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Cataloguing data can be found at the end of this publication.

Luxembourg: Office for Official Publications of the European Communities, 2000

ISBN 92-828-9023-6

Printed in Belgium

PRINTED ON WHITE CHLORINE-FREE PAPER

Frontispiece
Luskentyre Beach, looking out at the North Harris Hills, Isle of Harris, Outer Hebrides, western Scotland, United Kingdom, (Photographer: Jason Jordan).

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FOREWORD

This book is one of the outcomes of the expert workshop "Climate change and Sea level research in Europe: State of the art and future research needs" held at Mataro, Spain, in April 1997, organised under the auspices of the European Commission's Research Directorate General.

As the workshop title indicates its aim was to review the state of sea level research in Europe and to identify the most important needs for future research. After many years of sea-level research under the European Commission's Environment and Climate programmes it was thought timely to summarise the main achievements, to identify gaps and needs and to reinforce links to application and management. This is also consistent with the objectives of EC's Fifth Framework Programme, which stresses the need for multidisciplinary approaches and of the stronger involvement of users and stakeholders in the research effort.

The book reflects these new requirements and describes recent developments in the understanding and prediction of changes in sea level, its effects on the European coastline and its consequences for coastal zone management.

The latest scenarios from the Intergovernmental Panel on Climate Change (IPCC) indicate an increase in global mean sea level of some 18cm per century. The book discusses the concept of sea level change in response to global warming. Very importantly it stresses the need to consider local sea level changes, as sea levels will evolve differently in different parts of the world. Equally important, it underlines an integrative view of coastal processes and development, and stresses the need to view climate variability and change and coastal change as intricately linked through a number of processes of which sea level and storminess are of paramount importance.

We thank the editors and the contributors for all the dedication and hard work they have put into this book. It deserves a wide dissemination as a source of inspiration and an important reference for researchers and stakeholders in the field of climate and sea level change.

Anver Ghazi, Denis Peter and Ib Troen
European Commission, Biodiversity and Global Change unit, Brussels, Belgium.

PREFACE

The advancement of scientific knowledge involves the reduction of uncertainties and a search for more detailed description and explanation of phenomena; sea level research is no exception. A simple model of synchronous change in the level of the world's oceans in response to global climate change is no longer appropriate. Recognition of the complex nature of sea level change is reflected in the growing use of the term "relative sea level change". Spatial and temporal responses of the level of the sea surface and of the earth's crust are varied, as is the coastal geomorphological response. Arguably the major challenge facing the community of scientists working in the field of sea level change is to explain and estimate relative sea level and associated coastal changes at a local level, producing assessments of value to local communities on a decadal to centennial timescale.

The Mataro meeting examined the present status of sea level research in Europe, and evaluated the probable need for future work. This book begins with the recommendations drawn up at the meeting, followed by five chapters which describe recent work in the field. Chapter one considers our present understanding of sea level change on a global to regional scale, outlining some major issues and recent advances. Chapter two looks at sea surface variations, both temporally and spatially. Chapter three examines recent work on rapid change, involving secular variations, storm surges and tsunamis. Chapter four considers the effect of relative sea level change on coastal processes. Chapter five examines the implications for coastal zone management. Collectively, these chapters are intended to reflect the breadth of approaches used and results available at the present time. It is not claimed that this book is fully inclusive of all sea level research being undertaken in Europe, but rather it is a comprehensive summary of the key developments taking place.

Over thirty scientists contributed to this volume. The contributors and their affiliation are listed after this preface. All are thanked for their efforts in outlining results of aspects of their work. Additional thanks are due to Dr Anver Ghazi, Dr Ib. Troen and Mr Denis Peter of the European Commission, for their initiative in organising the Mataro meeting, and for their enthusiasm for the field of European sea level research.

A number of people were involved in assembling this book in a form suitable for publication, including at the Centre for Quaternary Science, Coventry University, Mrs Lucy Holloway (editorial assistant), together with Mrs Gillian West and Mrs Ann Daly (secretarial support), Miss Erica Milwain, (cartographer), Dr. Sue Dawson and Mr Jason Jordan. At the European Commission, Brussels, the help and technical advice of members of the Climatology and Natural Hazards Unit enabled the many logistical and technical problems in the production of this book to be overcome.

David E. Smith
Centre for Quaternary Science, Coventry University, United Kingdom.

RECOMMENDATIONS

ROBERT J.N. DEVOY, THOMAS A. DE GROOT, SASKIA JELGERSMA, IGINIO MARSON, ROBERT J. NICHOLLS, ROLAND PASKOFF & HANS PETER PLAG.

At the meeting in Mataro, Spain (April 1997) key recommendations were proposed as a focus for future research in sea level changes and coastal flooding, coastal processes and coastal zone management. Sections 1 and 2 present these recommendations and the rationale behind them.

1. SEA LEVEL CHANGES AND COASTAL FLOODING

Hans Peter Plag, Iginio Marson & Roland Paskoff

These recommendations identify the major sources of uncertainties in our understanding of past and present sea level variability and in predictions of future trends. They point out the research tasks required in order to achieve a reduction of these uncertainties. The participants emphasised the need for continual technical innovation in the observational techniques applied to the monitoring of sea level and related changes as well as the methods used to establish past sea level states and proxies of sea level change.

1 *Establish improved constraints on past and present changes in the mass and volume of the oceans for time scales of:*

- *centuries to several millennia based upon a combination of observational evidence (geological, geomorphological, geochemical and archaeological data) and geophysical models;*
- *years to several decades based upon the available tide gauge records taking into account all possible regional to local effects (such as crustal subsidence or uplift, changes in ocean circulation, effects of developments of river basins);*
- *days to years based upon satellite altimetry and related remote sensing data.*

To establish these constraints with the required accuracies, the development of improved techniques, for example in dating as well as satellite monitoring, is of key importance.

Concepts of a "global sea level curve" and particularly of a "global sea level rise" play a prominent role in the current climate change debate. Particularly for the assessment of future impacts of climate change and the planning of actions to mitigate these, a knowledge of variations in the pattern and timing of future relative sea level changes is of high social and economic value. The possibility of a significant sea level rise over the next 50 to 100 years has invoked considerable public interest.

Attempts have been made to identify a "climate signal" in global sea level, which can be expected for two reasons: (1) a change in the heat content of ocean water leads to a volume change (this effect has entered the discussion under the slogan "thermal expansion"); and (2) mass exchange between ocean and the cryosphere. A knowledge of the global ocean water mass and volume (GOMV) as function of time constitutes a crucial constraint for the reconstruction of past climates, as well as the validation of models used to predict future changes in the GOMV. However, due to the fact that the Earth is a deformable, gravitating body, there is no simple relation between sea level observed at any point on the Earth's surface and the GOMV. Up to now, establishing past variations in the GOMV has been hampered by the complexity of this relationship, and the available global curves and trends are unsatisfactory.

Global "sea level curves" for the last 20,000 years have been established from isotopic data taken as proxies for the amount of fluid water as well as geological data from sites considered to be representative for a "eustatic" sea level. Moreover, these curves have been used as input for geophysical sea level models and have been iteratively improved. Nevertheless, the current knowledge of past volume and particularly mass changes in the oceans due to the disintegration of the ice sheets of the last ice age is still unsatisfactory and the large uncertainties in the GOMV function over time constitute a serious limitation for our capabilities to predict future changes in that function.

For the last 100 to 150 years, sea level curves as well as global trends have been determined in several studies. However, all of these curves use basically the same global data set of monthly (or annual) mean sea level records obtained by coastal tide gauges. Despite the common database, the results display a considerable scatter indicating the degree of freedom in data preparation, selection, and corrections inherent both in the data set and the complex relation between relative sea level at the coast at a single tide gauge and the GOMV. Relative sea level at any given tide gauge may be affected by many causes operating on different spatial and temporal scales which separate into two broad classes, namely those changing the absolute sea level and those resulting in vertical land movements. Only a rigorous analysis of the global tide gauge data set making use of all available information concerning local effects at the tide gauge location can determine reliable uncertainty bounds for the global signal. Up to now, no such rigorous analysis is available. The uncertainties given for the global trends determined in the various studies are of the order of less than 1 mm/a but these uncertainties are pure statistical quantities. The uncertainties due to the complexity of the physical problem may in fact be much larger than the numbers given.

The presently available results from the Topex/Poseidon (TP) mission reveal a spatial variability in the trends over the total period of observation (three years) of the order of +/- 2 mm/a. In other words, TP sees a spatially varying decadal sea level variability which is of the same order as the total global sea level signal determined from the TP observations. A detailed study of the spatial variability in these short-term trends in relation to other parameters such as sea surface temperature and wind field is required to improve our understanding of these variations and the overall trends.

2 *Improve the knowledge of the current mass balance of ice sheets and glaciers as well as the prediction of the future mass balance by:*

- *extended measurements of accumulation and mass output of the ice sheets including an improvement of the observational technologies;*
- *improvements of dynamic models of the ice sheets with particular emphasis on the long time scales relevant to ice sheet dynamics;*
- *using high resolution forcing functions particularly from GCMs for mass balance calculations;*
- *extending studies to large glaciers to cover the broad spectrum of glaciers.*

A major contribution to GOMV changes in the past resulted from changes in the cryosphere, particularly in the large ice sheets. The uncertainties in our knowledge of the current mass changes in the cryosphere are too large to even constrain the sign of the total signal. A reliable prediction of the future GOMV changes or trends will only be possible if the uncertainties in our knowledge of the current mass changes in the cryosphere can be narrowed. Moreover, the long (centennial) time scales dominating the response of ice sheets to climate variations pose a serious challenge for dynamic models of ice sheets, and considerable improvements in the models are required particularly with respect to their ability to model their long-term behaviour. GCM studies and observational evidence suggest that climate change as well as the response of the ice sheets to this forcing are spatially highly variable. Currently, mass balance calculations for ice sheets are based on forcing functions averaged over large regions such as, for example, Greenland. It is widely accepted that this approach may be inappropriate and more detailed calculations with high resolution forcing functions are urgently needed.

Glaciers exhibit a broad spectrum in size, and the response of glaciers to climate change can be expected to be dependent on the size of the glaciers. Up to now, most studies have concentrated on smaller glaciers. To cover the broad spectrum of glaciers, special efforts should be made to include large glaciers in future studies.

3 Reduce the uncertainty in the prediction of thermal expansion of the oceans by:

- *obtaining further observations of three-dimensional temperature, salinity and heat flux into the ocean and temporal changes of these quantities;*
- *clarifying the cause of inter-model differences;*
- *the development and validation of advanced models.*

Besides mass exchange with the cryosphere, thermal expansion of the oceans is considered as a major contributor to future GOMV changes. However, the uncertainties associated with the estimates of the effect of thermal expansion on GOMV over the last 100 years are still very large. The uncertainties mainly result from a lack of sufficient observations of three-dimensional temperature and salinity in the oceans. Models used to estimate thermal expansion display considerable inter-model differences with the cause of these differences not being understood in depth. Moreover, the models exhibit a high degree of simplification. Consequently, in order to improve the reliability of predictions of the future effects of thermal expansion on GOMV more advanced models are required as well as additional observations to validate these models.

4 Estimate the contribution of the various components of the complete hydrological cycle to mass and volume changes of the oceans, including (besides thermal expansion and changes in the ice sheets and glaciers), for example variations in permafrost areas, extraction of groundwater and effects of irrigation, effects of development in river basins, reservoir building, soil sealing and deforestation.

To reduce the uncertainties in these estimates, improvements of the database related to the global hydrological cycle, as well as innovations in nearly all monitoring techniques applied to the various components of the cycle are key issues.

GOMV are characteristics of the oceans as reservoirs of the global hydrological cycle. Consequently, GOMV may be affected by changes in the other reservoirs of this cycle such as groundwater, soil moisture, humidity of the air, terrestrial surface water, ice sheets, glaciers, or permafrost. Studies of particular anthropogenic influences on these reservoirs and their potential contribution to changes in GOMV have demonstrated that some human interferences such as deforestation, groundwater extraction, irrigation, river regulation or soil sealing may significantly contribute to GOMV changes. However, there are considerable differences between estimates obtained in the various studies basically because they were based on insufficient data. Therefore, more thorough studies of the fluxes and reservoir changes in the hydrological cycle based on the broadest possible database are necessary to clarify the contribution of the various processes to GOMV changes.

5 Improve the models for prediction of future volume and mass changes of the oceans, and the devlopment and integration of Recommendations 1 - 4.

Current models used for the prediction of future changes in GOMV in the context of, for example, the IPCC assessments are still based on many simplifications, which currently may be justified by the considerable uncertainties in the quantification of the basic processes affecting the GOMV. However, as these uncertainties can be expected to be reduced in the future, efforts should be made to improve the predictive models particularly by incorporating more feedbacks and by setting up models with at least two spatial dimensions.

6 ***Continue, improve and extend the on-going integrated programme of global sea level monitoring including:***

- ***tide gauges at adequate spatial density along coastlines to monitor relative sea level changes;***
- ***stations with combined tide gauges and GPS (and DORIS) receivers at selected sites for altitude calibration;***
- ***GPS and absolute gravity observations at tide gauges and elsewhere for the determination of rates of crustal movement;***
- ***a programme of TOPEX-class altimetry for near-global monitoring of geocentric sea level;***
- ***effective mechanisms for free exchange of data of all types (especially within Europe).***

Observations of sea level variations on different time scales are highly relevant as they, on the one hand, allow for the analysis and description of sea level variability itself as a prerequisite to a better understanding of the causes and, on the other hand, constitute crucial constraints for models related to sea level such as hydrodynamic models or coupled atmosphere-ocean circulation models.

Tide gauges have been operated at coastal sites since the beginning of the nineteenth century. Today, more than a thousand tide gauges are operating globally. The data set of monthly mean values computed from the tide gauge records and collected at the PSMSL constitutes one of the most valuable climatological data sets. In particular, the long records with spans of several decades are a unique source of knowledge for the interannual to centennial sea level variability, which is directly related to climate variability. It is highly recommended not to let this data set be interrupted. In order to maintain or even improve the spatial coverage of coastal tide gauges, it is also required to continue to operate tide gauges with shorter records or even set up new tide gauges in areas up to now not having any recordings. GLOSS has put forward an implementation plan which supports this by giving exact locations for a global tide gauge network fulfilling both the temporal and spatial aspect.

Satellite altimetry has turned out to be a most valuable source of information on intra- to interseasonal sea level variability with a near global coverage. Current and future missions strongly rely on tide gauge calibration. For this purpose, suitable tide gauges have to be co-located with space-geodetic equipment such as GPS or DORIS receivers, which provide a high-accuracy control of the tide gauge benchmark movements. Here again the GLOSS implementation plan includes a list of suitable stations as well as recommendations for the observational strategies. GLOSS also encourages improvements in the density of the tide gauge network and has accepted proposals for EUROGLOSS and MEDGLOSS networks (including GPS, absolute gravity, etc.). It is strongly recommended that the implementation of GLOSS be supported both at national and international level.

Tide gauges measure sea level relative to a benchmark on land. Separation of crustal movements from the "oceanic" contribution to relative sea level changes depends upon the accuracy to which the vertical movement of a benchmark is known. Space-geodetic observations at the tide gauge provide the best available means to observe benchmark movement with sufficient accuracy. Following the recommendation of respective international working groups, tide gauges to be used for absolute sea level observations should be co-located either permanently or episodically with GPS receivers and absolute gravity measurements should also be made wherever possible.

Topex/Poseidon has revolutionised our knowledge of the spatial variability of sea level on intra- to interseasonal time scales and improved our understanding of the forces producing this variability. At decadal and longer time scales, coupled atmosphere-ocean phenomena contribute dominantly to the climate variability. Therefore, extending these sea level observations to decadal time scales would provide an invaluable key to a better understanding and eventually to a long-term prediction of climate variability at interannual to decadal time scales. Thus, a continuous programme of Topex-class altimetry of the global sea levels will provide observations crucial to understanding the forcing of climate variability at interannual and decadal time scales.

To promote to the largest possible extent studies of sea level variability and its relation to climate variability it is required to guarantee free access to relevant data archives for all interested scientists. Relevant archives include meteorological and climatological databases, oceanographic data, crustal movement observations particularly those acquired with space-geodetic techniques, monthly mean sea levels or hourly values derived from tide gauge observations, and satellite altimetry observations or other remote sensing data of the ocean.

7 ***Set up and validate models of contemporary crustal movements covering global to local (e.g. estuary and delta) scales. Evaluate the impact of reference systems and the geoid as well as their temporal variability.***

In many observations of global change phenomena including relative sea level changes and changes in ice mass, crustal movements due to tectonic, anthropogenic or other causes are obscuring the global change signal that is being sought. Therefore, absolute gravity and GPS or other space-geodetic observations are required particularly at tide gauges but also at other locations in order to get a clear picture of crustal motion. It should be pointed out that combined absolute gravity and GPS observations give valuable information on the underlying mechanisms. Thus, these observations provide constraints necessary for the validation of global and regional geophysical models of present-day crustal movements. Such models are required for decontamination of observations in order to separate relevant global change signals from other disturbing influences.

8 ***Assess the existing sea level and sea level related database in order to describe the data quality as well as to identify observational gaps both in space and time. An example of such a gap is the missing link between geological and instrumental data, where technical improvements are required in order to close this gap.***

In the past few decades, several databases with sea level and sea level related observations have been compiled. For example, the PSMSL has over the last more than 60 years built up a database of global monthly mean sea levels and thus compiled one of the most valuable climatological databases. Several research groups have compiled regional or global databases of relative sea level changes over the last 20 kyrs, comprising various types of samples used to derive the timing and location of former sea levels. Oceanographic data sets have been compiled for climatology as well as temporal variability of, for example, sea temperature and salinity. Revised analysis of meteorological observations of the past several decades have been carried out and have resulted in data sets providing information on the atmospheric forcing of sea level variations. However, the available data sets are of varying quality and all have their specific advantages and disadvantages. Thorough assessments of the existing data sets are urgently needed in order to assess the data quality based on a prescribed, as far as possible homogeneous, and structured method. Assessing the data quality as well as the applicability and limitations of the various data sets for global change studies would help to ensure an optimal use of the available data and to avoid misinterpretations or waste of the available resources. Moreover, the identification of observational gaps both in space and time would help to direct scarce research resources to those areas with the highest potential for furthering our knowledge and understanding.

9 ***Study sea level variability in relation to atmospheric forcing from time scales of hours (storm surges) to decades.***

Potential hazards for low-lying coastal areas not only arise from a possible future rise in sea level but also will result from a change in the sea level variability due to variations in the atmospheric forcing or the response of the ocean to the forcing. Therefore, studies of sea level variability in relation to a changing atmospheric forcing are required on time scales relevant to storm surges, where both an increasing frequency as well as severity may have serious effects on the anthroposphere. At longer time scales of up to several decades, coupled atmosphere-ocean phenomena are dominantly influencing climate variability, and studies of atmosphere-ocean interactions at these time scales may provide the urgently needed basis for understanding how future climate changes may affect climate variability on seasonal to centennial time scales.

10 ***Establish the means for downscaling of a predicted global sea volume signal to local and regional scale on the basis of geophysical models of the atmosphere, ocean and solid Earth in order to make progress towards a prediction of sea level changes at local scales.***

It must strongly be emphasised that future changes in GOMV as discussed in, for example, the IPCC assessments cannot be directly equated to changes in relative sea level at a given location. Neither sea level changes due to mass exchange between ocean and cryosphere or terrestrial hydrosphere nor those due to volume changes as a consequence of an increasing heat content of the ocean can be expected to be globally uniform. Mass exchanges between ice and ocean result in deformation of the Earth as well as changes in the geoid both affecting the distribution of ocean water and thus relative sea level. Geophysical models can be used to compute the water distribution and relative sea level if the origin of the mass added to the ocean or the destination of the mass taken from the ocean (i.e. the ice load history) is known. Models of heat transfer between atmosphere and ocean under changing climate conditions can be developed and used to predict spatial variability in the thermal expansion of the ocean waters. General circulation models of the atmosphere provide predictions of future changes in the prevailing wind fields. In combination, these models can be used to accomplish a downscaling of the predicted GOMV signal to regions or even local areas and thus provide a sound basis for an integrated coastal zone management particularly in areas with economic and social consequences arising from climate and sea level change . Directly equating future changes in GOMV to regional or local relative sea level changes, as is still often the case even on national level, may feed considerable economic resources into mitigation attempts in areas of relatively low risk while in other areas a much higher risk may not properly be recognised and timely attempts to mitigate the consequences may not be made.

2. COASTAL PROCESSES AND COASTAL ZONE MANAGEMENT

Saskia Jelgersma, Robert J.N. Devoy & Thomas A. De Groot

2.1 Coastal processes

Whilst present understanding of the processes operating in the coastal zone has improved dramatically, the importance of long-term influences such as sea level rise require more study before the practical relevance of such changes to coastal communities can be quantitatively assessed. The need to determine the detailed impact of climate change on coastal processes forms the basis of the recommendations listed below. These are as follows:

1 ***Improved understanding of the role of the sediment budget, since this is critical in determining shoreline location and the survival of coastal wetlands and intertidal habitats as well as dune areas.***

On many coasts, sediment availability has fallen in the late Holocene as a result of both human and natural causes. In some areas this has led to serious erosion along coastlines and consequent effects upon coastal communities. Combined with relative sea level rise, this alteration in sediment availability has particularly serious effects in some areas. There is a need to better understand long shore and cross shore supply and transport at a range of scales, including episodic extreme events, as well as examine the importance of sediment sinks. In this connection, there is also a need to study the role of the shoreface as a source of such sediment. The input of sediments from river systems to the coast needs to be examined, both in terms of bed load and suspended load and the effect of shore erosion upon sediment delivery needs also to be studied. The importance of understanding the erosion and accretion of sediments in coastal wetlands and associated intertidal areas needs to be stressed in the context of sea level rise.

2 Development of robust coastal typologies of European coasts for modelling and management purposes.

Whilst coastal types such as beaches, wetlands, deltas can be recognised, more detailed division of coastal types is necessary as a framework to model coastal response to sea level rise and other changes (see 3 below). This activity is recognised by the IGBP-LOICZ programme. It will also provide an improved basis for the transfer of scientific knowledge to coastal zone management. Particular attention should be focused upon sensitive coastal areas such as deltas, estuaries and areas of coastal wetlands where changes in the climate in terms of storminess and sea level rise combined with land movement may be expected to have particular effects.

3 Improve models of coastal development, to include the effects of increases in sea level rise and changes in sediment budgets.

Present models of coastal development have been devised for present conditions and will perform poorly under changing conditions. In many areas of the European Coastline, increases in relative sea level rise require further development of such models. The complexity of factors affecting coastal change are such that broad, conceptual/behavioural models need to be developed in parallel with predictive models for management purposes.

4 An assessment of the impact of extreme and unusual events, particularly tsunamis and storm surges, on sensitive coastal areas.

Most knowledge of sea level change relates to gradual change, but in the past and possibly in the future more rapid change have taken place. Studies of the coastal response to rapid sea level change, including possible accelerations in sea level rise, are needed.

Changes in the frequency and magnitude of storm surges appear to be a consequence of climate variability and change. The response of the coastline to variations in the intensity and frequency of storms needs to be assessed, notably with a view to determining the significance of thresholds. It is now generally recognised that tsunamis occur frequently along european coasts , including both the epicontinental Mediterranean and North seas as well as along the Atlantic coast. Their effects as agents for coastal change, especially in the breaching of thresholds and their impacts, particularly along vulnerable coasts, need to be assessed. Studies of the impacts of individual events and of long term trends including both tsunamis and storm surges are needed.

2.2 Coastal Zone Management

Robert Nicholls

The coastal zone is (i) a series of interacting sub-systems with uncertain boundaries , (ii) a system with a complex and poorly understood response to external forcing such as global sea level rise, and

(iii) a system which is increasingly affected by anthropogenic use and management. Effective management of the coast therefore requires integration, development and application of knowledge from a wide range of disciplines including coastal scientists (morphodynamics, hydrodynamics, geology), ecologists, and socio-economists (including the policy makers). Despite efforts to develop integrated coastal zone management (ICZM) the reality is that few national entities are applying an ICZM due to a strong bottom up approach to the problem, financial and cultural implications, conservative applications of legislation, limitations of existing CZM models and government structures.

Within these boundary conditions, the objectives of ICZM can be defined as follows :

- Sustainability of the responses to demands on the coastal zone;
- Shoreface planning integrated into the coastal zone framework;
- Balancing of competing demand;
- Cohesion of decision making;
- Flexibility of responses to coastal problems.

The following recommendations will facilitate ICZM, but are developed largely from a coastal science perspective, and reflect attendance of the workshop.

1 *Coastal processes and their understanding should play a key role in ICZM. They form the factual basis for coastal zone evolution.*

2 *A key issue is the understanding of the coastal processes within defined coastal cells that form the spatial (and often temporal) basis for coastal science analysis and units for coastal management. Long-term (continuous) multidisciplinary monitoring of selected coastal cells is required for a range of coastal types, with or without protection measures applied, in order to create feedback and understanding of the processes involved. This could be done by the establishment of (low cost) coastal observatories in selected areas over a two to three year period, later to be upgraded to permanent sites on a regional basis. Monitoring will provide the data for assessments of coastal sensitivity/susceptibility and resilience, although these are difficult to measure, especially where extreme events are concerned.*

3 *The availability and accessibility of data is a key constraint to ICZM. Better, more complete and more flexible databases of the full spectrum of coastal data is needed to allow an integrated analysis of coastal changes in defined coastal cells.*

4 *The process of integration requires investigation and conceptual development. This includes improved integration between physical, biological and social sciences as well as policy making.*

5 *Exiting assessments of climate change and global sea level rise impacts have focused on the extreme human response of "do nothing" and "protect everywhere". More detailed assessments of the full range of possible response options are required.*

6 *Attention should be paid to the linkage of existing expertise with the skills and needs of the end-user community. To ensure this, existing funding programmes (e.g. LIFE and LEADER), besides 'traditional' ones like MAST or ENVIRONMENT & CLIMATE, should be used and expanded to increase the flow of current expertise towards the community/user level.*

1
GLOBAL CHANGES IN THE VOLUME AND MASS OF THE OCEAN

SARAH C. B. RAPER, *Chapter editor*

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1.1 OVERVIEW

In this section, climate related changes in the volume and mass of the oceans are considered. The concern here is with possible natural and/or anthropogenic changes in climate and the effect these will have on the ocean volume and mass over the next 100 years. Although, as will become evident in this and the following section on regional changes, sea level changes have been and are likely to be highly spatially heterogeneous it has been customary to express changes in the volume and mass of the ocean in terms of equivalent global mean sea level change. However, the fact that the ocean volume and mass related global mean sea level change is a hypothetical quantity and only one of many factors which will affect sea level at a particular location should never be lost sight of.

Although we are primarily concerned with changes over the next 100 years it is necessary to study the past on different time scales. Clearly we need to study in detail the last 100 years when observations are the most plentiful, to develop our understanding of the physical processes and to develop our models. However, the oceans and ice sheets are so vast that they are still reacting to climate changes which occurred in the more distant past. Because of the long response times and also because a knowledge of the past provides the key to the future, ocean volume and mass changes from the late glacial to the present are also of great interest.

In section 1.2 we present a selected review of observation based evidence for past changes in the ocean volume and mass, starting with the more distant past which relies on geological evidence, then looking at the evidence from tide gauges and finally from satellite altimetry. Section 1.3 considers the factors which contribute to changes in the ocean volume and mass. The main factors which have been considered are thermal expansion and the melting of land based ice. The latter is treated in three categories, first glaciers and ice caps, second the Greenland Ice Sheet, and third the Antarctic Ice Sheet. In section 1.4 we bring together some conclusions about the estimated changes over the past and the modelled contributions. The uncertainties in the projections and how those projections should be used is considered.

1.2 OBSERVATION BASED ESTIMATES OF OCEAN VOLUME CHANGES INTRODUCTION

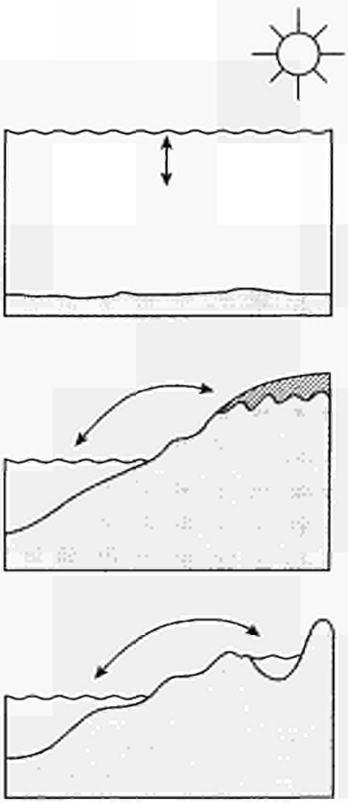
1.2.1 Introduction

First a contemporary view of the attempts which have been made to piece together a picture of past ocean volume changes is presented. For the more recent past, the problems in estimating global mean sea level change from tide gauge data arise from the spacio-temporal distribution of the data and from vertical land movements. The methods used to try to overcome these problems are discussed as well as planned improvements in future monitoring. These plans include the use of satellite altimetry data which are already providing complete ocean coverage for recent years.

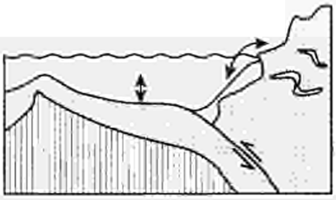
1.2.2 Ocean volume and sea level changes from geological evidence

Michael Tooley, Alastair Dawson & Antony Long

1.2.2.1 Introduction. The volume of the ocean is approximately 1322 x 10^6 km^3, and its total mass is estimated to be 143 x 0^6 metric tons (Viglieri, 1966). The volume of the oceans appears to have changed little on timescales of 10^8 to 10^9 years. However on shorter timescales (e.g. 10^7 years) it may fluctuate by up to 6%, largely as a result of the growth and decay of ice masses (Fairbridge, 1987) adding or subtracting up to 80 x 10^6 km^3 of water to or from the world's oceans. In addition to ice volume changes there are other mechanisms (Figure 1.1) that have affected the volume of the ocean and these are steric or thermohaline alteration, the recharge and depletion of ground water including permafrost, and crustal deformation (Plint *et al.*, 1992). These mechanisms operating together in unison or alone can alter the position of sea level by a few millimetres to over a hundred metres on timescales of seconds to millions of years, but the position will not be the same over the world.



MECHANISM	Time Scale (years)	Order of Magnitude (m)
1) Ocean Steric (thermohaline) Volume Changes		
Shallow (0-500m)	0.1-100	0.01-1
Deep (500-4000m)	10-10 000	0.01-10
2) Glacial Accretion and Wastage		
Mountain Glaciers	10-100	0.1-1
Greenland Ice Sheet	100-100 000	0.1-10
East Antarctic Ice Sheet	1000-100 000	10-100
West Antarctic Ice Sheet	100-10 000	1-10
3) Liquid Water on Land		
Groundwater Aquifers	100-100 000	0.1-10
Lakes and Reservoirs	100-100 000	0.01-0.1



4) Crustal Deformation

Lithosphere Formation and Subduction	100 000-10^8	1-100
Glacial Isostatic Rebound	100-10 000	0.1-10
Continental Collision	100 000-10^8	10-100
Sea Floor and Continental Epirogeny	100 000-10^8	10-100
Sedimentation	10 000-10^8	1-100

FIGURE 1.1 Mechanisms affecting the amount of sea level change and ocean volumes on a range of time scales (based on Reuelle 1990, in Plint et al., 1992).

It was a long held view that changes of sea level were global and were of the same magnitude and sign. They were called eustatic changes and were first defined by Suess in 1888 (see discussion in Tooley, 1993). However, the conclusion of International Geological Correlation Project (IGCP) 61 was that there was no globally valid sea level curve (Tooley, 1987): this conclusion had been anticipated by Morner (1976), who had argued that the existence of the geoid - the equipotential surface of the ocean set up by the opposing forces of gravity and the centrifugal force of the Earth's rotation - resulted in non-uniform but regionally coherent movements of sea level.

The recognition that there is no such thing as a eustatic sea level curve presents problems for those seeking to interpret data on the magnitude and patterns of Late Quaternary and pre-Quaternary sea level change. For example, the surrogate record of global ice volume changes derived from oxygen isotope studies of benthic foraminifera may be more properly considered as an approximation of long-term ocean volume changes rather than of sea level changes. Similarly, the several very detailed investigations of emerged and submerged coral sequences provide a more valuable record of ocean volume changes than of former changes in sea level. In this paper a review is presented of long-term sea level changes, geoidal changes and the interrelationships between reconstructions of Late Quaternary ice and ocean volumes, sea level changes and the earth rheological models that make use of such data. For the Late Quaternary the discussion is presented within the context of dating uncertainties that constrain the respective palaeoclimatic chronologies.

1.2.2.2 Long term changes of sea level. In the geological literature, unconformities in rock sequences have been explained as the consequence of eustatic sea level changes (Figure 1.2) and used as a basis for correlation from one basin to another both within and between continents. In order to aid hydrocarbon exploration of sedimentary rock sequences using seismic stratigraphy, outcrop sections and well log data in a multidisciplinary approach now defined as sequence stratigraphy (Plint *et al.*, 1992). Patterns of onlap, downlap, truncation and basinward shifts of coastal onlap are used to infer relative sea level changes: sea level rise is inferred from coastal onlap (cf. transgressive overlap in Tooley, 1982, for the Holocene). The magnitude of sea level rise is calculated by the amount of coastal encroachment and aggradation (Hallam, 1981). A series of sea level cycles has been recognised for the Phanerozoic. First order cycles lasting 200-400my are related to supercontinent break-up, spreading ocean ridges and consequential lowering of sea level. Second order cycles last 10-100my and as correlations have been established between several continents, eustatic sea level changes have been involved as an explanatory mechanism.

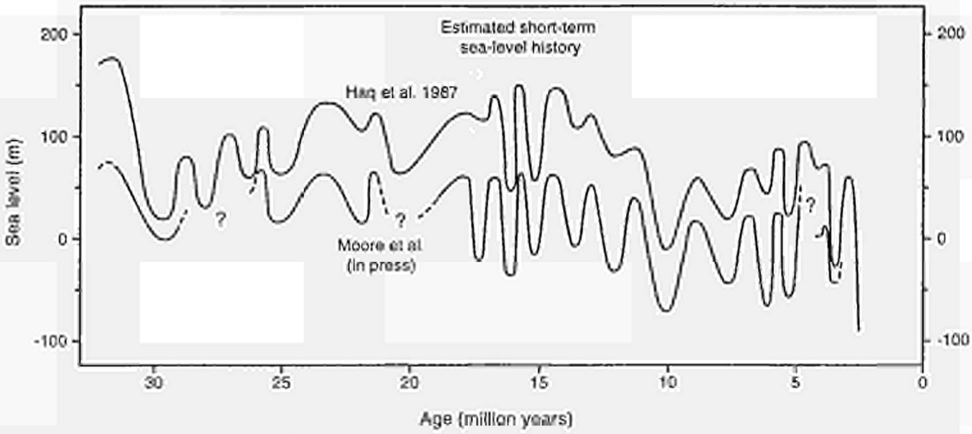


FIGURE 1.2 Sea level changes for the past 30My (Haq et al., 1987 and Moore et al., in Kerr, 1987).

Third order cycles (Figure 1.3) last from 1-10my and have been explained by the growth and wastage of ice masses, but continental correlation is poor and other mechanisms may operate.

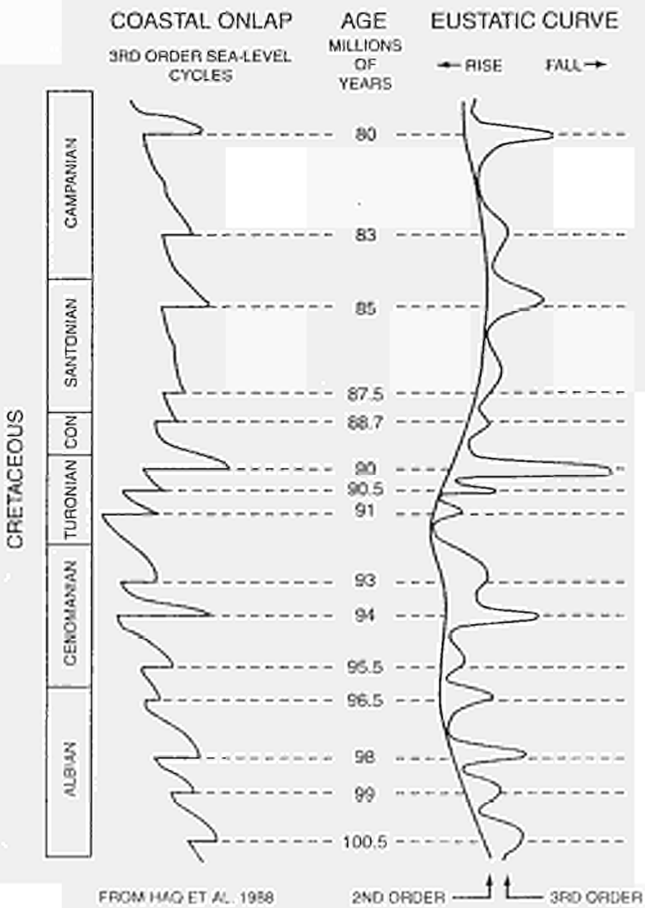


FIGURE 1.3 Third order sea level cycles and "eustatic" sea level curves for part of the Cretaceous (Haq et al., 1988 in Plint et al., 1992) when atmospheric CO_2 values were declining (see figure 1.4).

For example, Kauffman (1984 in Plint *et al.*, 1992) described third order marine transgressions in Cretaceous strata in western USA as a consequence of Cordilleran tectonism and volcanism and regressions by quiescence. Fourth (200-500ka) and fifth order (200-10ka) cyclicities have been identified from shallow water rocks of marine and pelagic origin, and are interpreted as the consequences of climatic changes arising from astronomical or Milankovitch Cycles (Plint *et al.*, 1992; Dawson, 1992). However, Berner (1994) concluded that the atmospheric CO_2 greenhouse effect has been the major control of climate during the Phanerozoic. He has demonstrated using a revised model of atmospheric CO_2 (GEOCARB II), that CO_2 values were more than ten times greater in the Lower Palaeozoic and early part of the Upper Palaeozoic, whereas during the later Upper Palaeozoic (the Permo-Carboniferous) and Cainozoic, values were low. These low values (Figure 1.4) were associated not only with glaciations (Berner, 1994), but also with the build-up of terrestrial vegetation (Chaloner and McElwain, 1997). The high values of atmospheric CO_2 are considerably greater than the post-industrial CO_2 rise and the highest stabilisation values projected to AD2500 (Wigley, 1995), and need to be correlated both with temperature and sea level fluctuations, as shown for the past 160ka (in Tooley, 1993, and Figure 1.4).

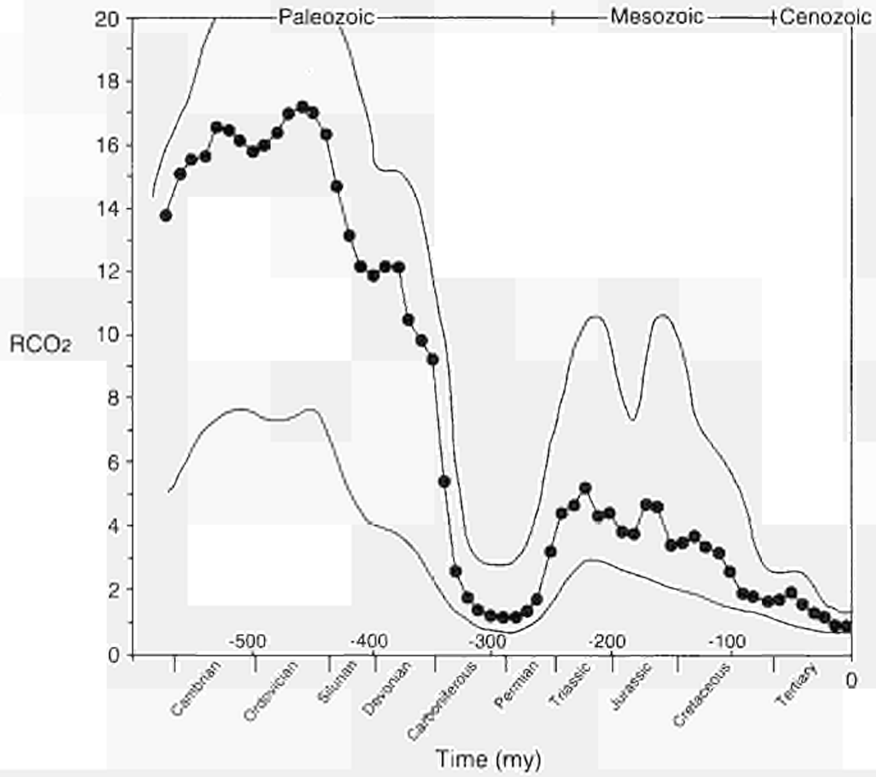


FIGURE 1.4 Phanerozic atmospheric CO_2 curve (Berner 1994). The vertical axis shows the ratio (RCO_2) of the palaeo-CO_2 level to that of the pre-industrial CO_2 level. RCO_2 is a ratio based on an estimate of the mass of atmospheric CO_2 at the time t divided by the "present" mass of atmospheric CO_2 (300ppmw - the pre-industrial level).

Plint *et al.* (1992) concluded that geoidal eustasy, as defined by Morner (1976), is unlikely to produce significant changes in relative sea level. At different timescales during the Phanerozoic, and particularly during the Late Quaternary it is inconceivable that the migration of the geoid has not occurred as a result of episodes of mountain building, isostatic recovery, the accumulation and decay of ice masses. Geoid deformation occurred during the existence of the Fenno-Scandinavian ice sheet. During the Younger Dryas Chronozone, Eronen (1987) had suggested that the geoid rose 38.5m at the centre of the ice sheet and 14.5m at the ice margin on the Norwegian coast; falling by similar amounts when the ice disappeared. Concomitant changes in the density of the mantle, crust and asthenosphere would have an instantaneous effect on the configuration of the geoid. The distribution of the geoid relief (Figure 1.5) shows a range of altitudes of -104m in Southern India and +74m over Irian Jaya and Papua New Guinea. This range, measured in relation to the best fit ellipsoid of the Earth, is similar to the difference between low relative sea levels of the last glacial age and the high relative sea levels of this interglacial age (Tooley, 1978). The distribution of geoidal lows and highs is meridional and is associated with the global pattern of positive and negative gravity anomalies (Fifield, 1984; Lerch *et al.,* 1994). Positive gravity anomalies are associated with geoidal highs and negative anomalies with geoidal lows. The negative anomalies over northern Canada and Fennoscandia are associated with areas of former glaciation: they are deficient in mass as the result of uplift, and gravitational equilibrium will return when relaxation is complete and the crust returns to its shape and density prior to glaciation. In the case of Fennoscandia there is a further 100 metres uplift to occur before this equilibrium state is reached affecting the regional pattern of geoid contours.

The abstraction and return of water from and to the world's ocean basins during a glacial-interglacial cycle will have an effect on the shape of the geoid as well as on sea levels through hydro-isostatic, geodetic, steric and earth rotational processes. Walcott (1972) described changes in the global distribution of sea level during deglaciation and identified three regions within which the behaviour of sea level varied significantly during the last 18ka. The first region was one in which rapid uplift and hence relative sea level fall occurred following deglaciation: the second region was one in which submergence occurred in the area immediately outside the glaciated limits: and the third region remote from ice loading in which coastal tilting results in emergence on continental shorelines because the rise of the continents relative to sea level is greater than the fall of the ocean floor due to enhanced water load. Clark *et al.* (1978) supported Walcott's conclusions but increased the number of regions with a characteristic sea level response following deglaciation to six (Figure 1.6).

To a certain extent empirical data support these models (for example Nunn, 1986; Martin *et al.*, 1985). Clark and Primus (1987) applied this methodology to predict sea level changes consequent upon a partial melting of the Greenland and Antarctic ice caps to AD2100 (Figure 1.7). They showed that observations of sea level change would vary from -200% to +120%. These results concur with earlier work on sea level change and earth rheological models which show that the attenuation of ice sheets does not result in a globally uniform rise of sea level. Indeed, simultaneously there can be both rises and falls of sea level in different parts of the Earth. This vindicates Morner's (1976) argument for regional eustatic studies. They are applicable not only to the Quaternary but also to earlier geoidal periods.

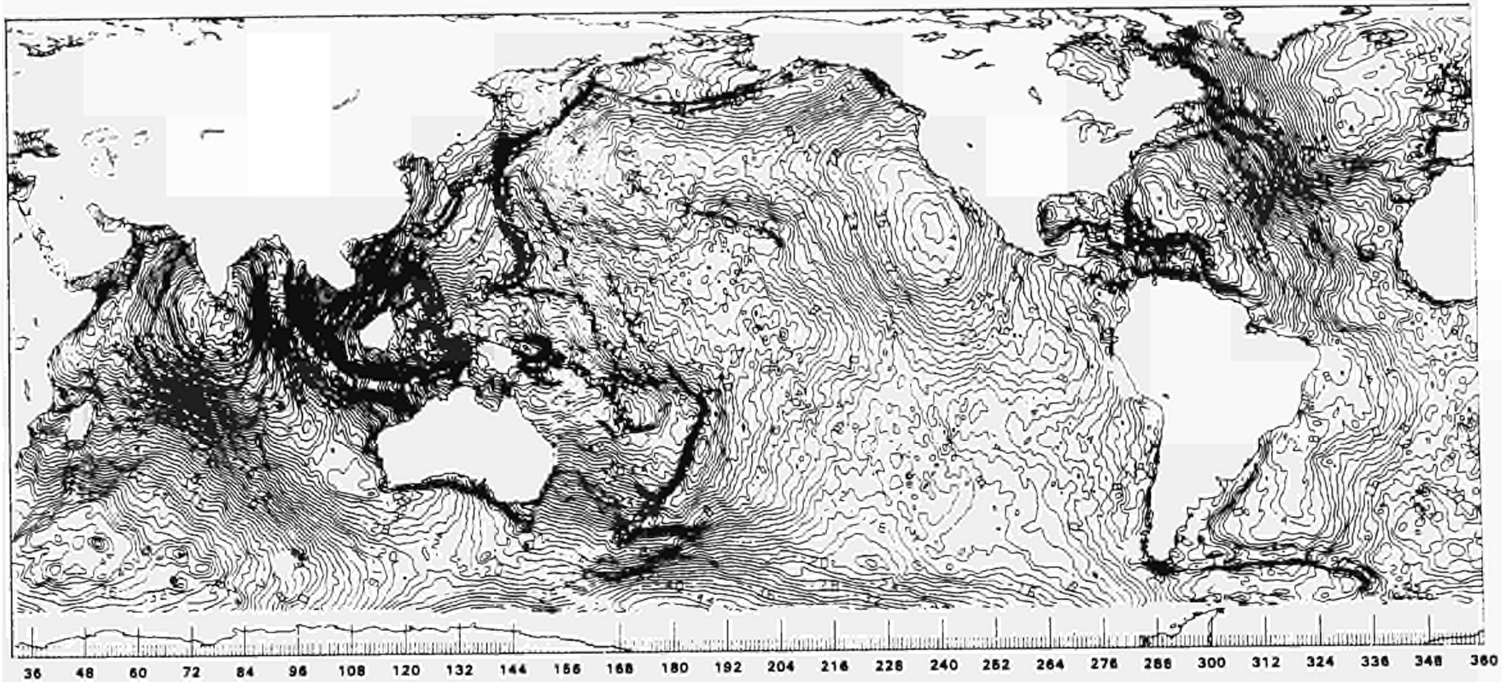
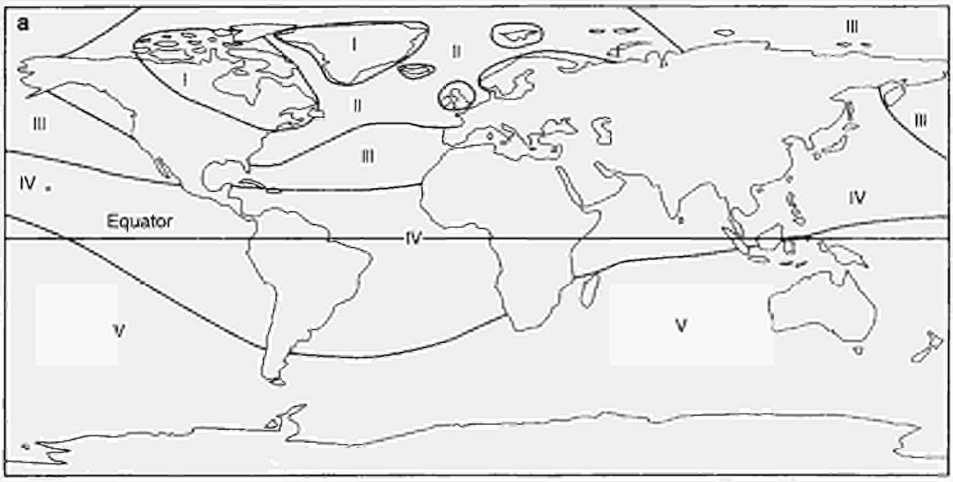


FIGURE 1.5 The "topography" of the geoid surface derived from the altimeter data of the NASA SEASAT mission in 1978. The long wavelength components of the geoid show undulations with a maximum amplitude of N180m (Marsh and Martin 1982).



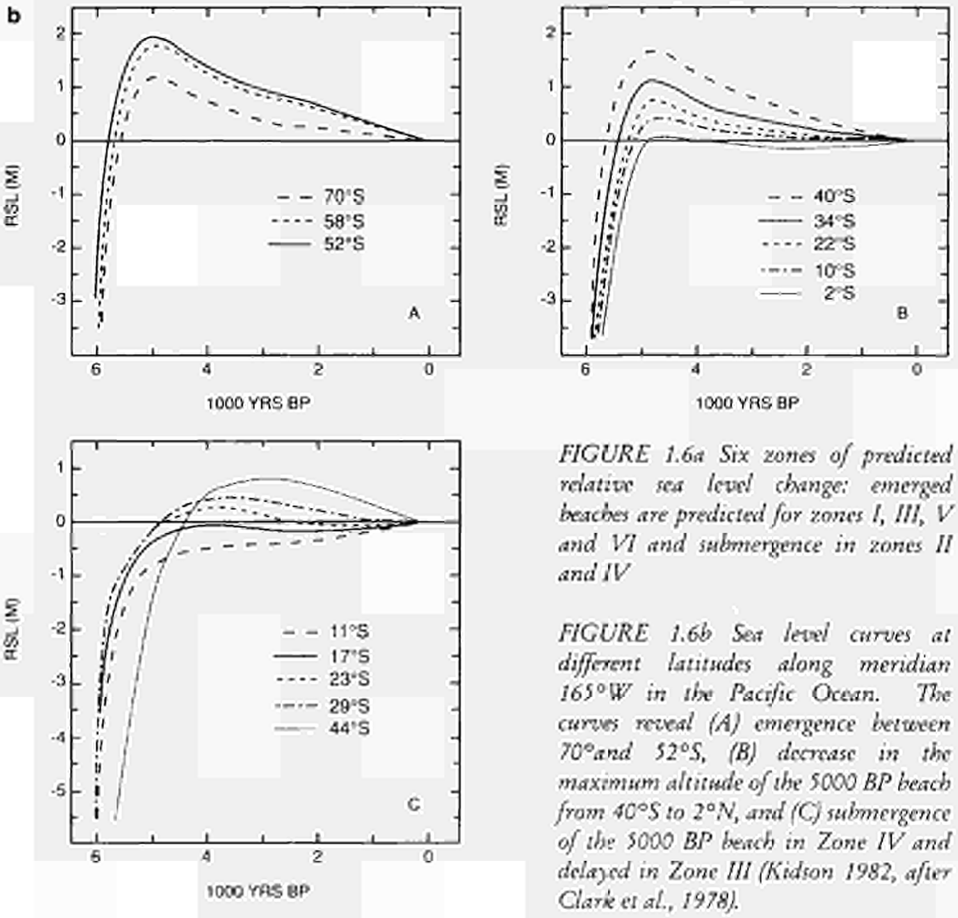


FIGURE 1.6a Six zones of predicted relative sea level change: emerged beaches are predicted for zones I, III, V and VI and submergence in zones II and IV

FIGURE 1.6b Sea level curves at different latitudes along meridian 165°W in the Pacific Ocean. The curves reveal (A) emergence between 70°and 52°S, (B) decrease in the maximum altitude of the 5000 BP beach from 40°S to 2°N, and (C) submergence of the 5000 BP beach in Zone IV and delayed in Zone III (Kidson 1982, after Clark et al., 1978).

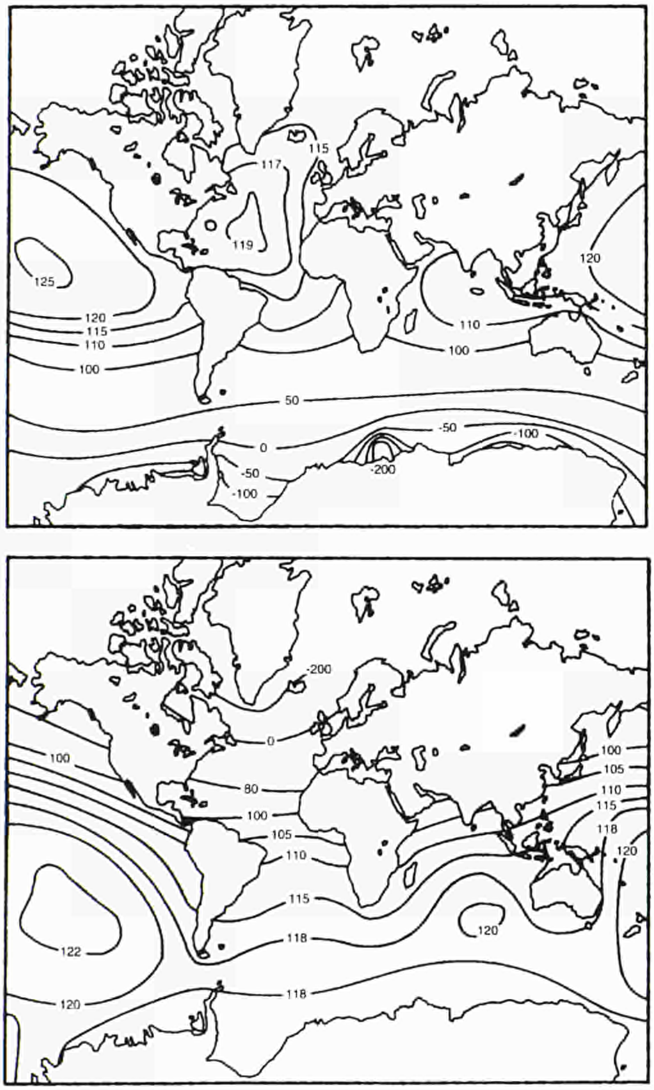


FIGURE 1.7 Predicted amounts of sea level change resulting from the melting of part of the Antarctic (top) and Greenland (bottom) ice sheets. Contours are in centimetres if the "eustatic" sea level rise is 100cm (Clark and Primus 1987).

1.2.2.3 Ocean volume changes. Total ocean volume is partitioned unequally between the world's oceans: the Pacific accounts for over 707×10^6 km^3, the Atlantic over 323×10^6 and the Indian Ocean 291×10^6 (Viglieri, 1966). Changes in the volumes of these oceans were characteristic over long geological timescales, and even during the Quaternary the estimated volume fluctuations of 6% may have been greater in one basin than another because of variations of gravity anomalies and hence geoid topography. Furthermore, variations in the temperature and salinity of deep and intermediate water masses will affect the volume of these water masses, the space occupied in the ocean basins and hence the height of the water column.

Attempts to reconstruct former ocean volumes are of value to our understanding of the magnitude and timing of palaeoclimatic changes for a variety of reasons. First, estimates of ocean volume derived from investigations of Late Quaternary oxygen isotope stratigraphy based on planktonic and benthic foraminifera (Figure 1.8) provide information on the growth and decay of Late Quaternary ice sheets (e.g. Shackleton, 1987).

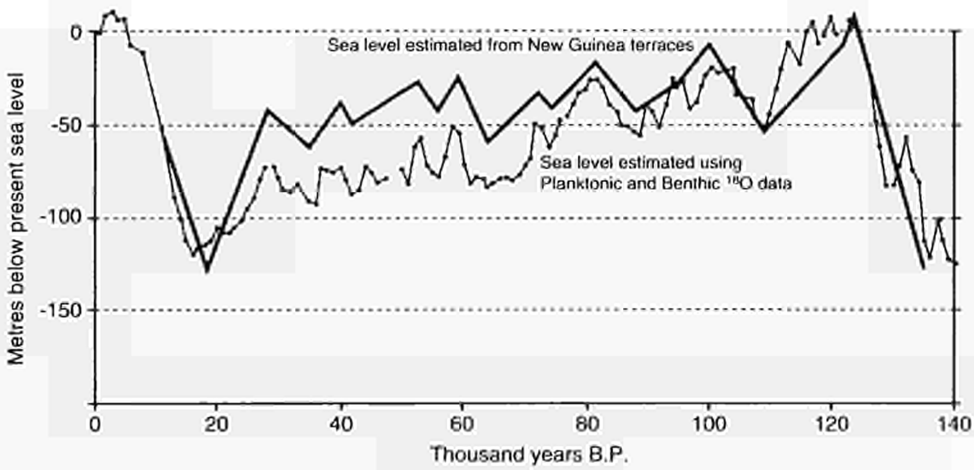


FIGURE 1.8 Oxygen isotope derived global sea level history compared with sea level data from Papua New Guinea (after Shackleton 1987).

These ocean volume changes have been used to constrain the timing and magnitude of ice sheet growth and decay but do not provide any regional information on ice sheet histories. The most accurate oxygen isotope records of ocean volume change are based on stratigraphical records of benthic foraminifera that provide a valuable chronology of the changes in the isotopic variation of seawater through time (cf. Shackleton, 1987).

Second, these estimates of former ocean volume change may be correlated, in turn, with global changes in palaeo-oceanography, ice sheet dynamics and climate (e.g. Fairbanks, 1989; Bard *et al.*, 1990, 1996; Blanchon and Shaw, 1995). Third, estimates of changes in ocean volume are required to constrain models of glacio-hydro-isostatic rebound at global to local scales (e.g. Peltier, 1994, 1996a & b; Lambeck *et al.*, 1990; Lambeck, 1993a, b& c, 1995a). In this regard, particular importance concerns estimates of the position of relative, regional sea level during the Last Glacial Maximum (LGM). Changes in ocean volume record net fluxes in the global hydrological cycle. Amongst others, this parameter includes fluxes from the ice sheets and glaciers to the oceans, thermal changes in ocean volume, as well as changes in groundwater palaeohydrology, particularly of lakes and wetlands. Some parameters have scarcely been investigated. For example, permafrost covers approximately 25% of the Earth's land surface yet the nature of permafrost degradation since the LGM is not known. It has been estimated that the area of continuous and discontinuous permafrost amounts to 24.9 x 10^6 km^2 which is equivalent to 25cm global sea level rise (Barry, 1985). However, no estimates have been given for stages of degradation of permafrost and the successive contributions to sea level rise during these stages.

1.2.2.4 **Measurement of ocean volume change.** Ocean volume changes (as opposed to sea level change) can be measured in two main ways: by estimating volumetric contributions from ice sheets and, indirectly, by measuring changes in the isotopic composition of ocean floor sediments. Two other approaches involve the measurement of sea level change using the geological record (e.g. analysis of former shoreline data, transgressive and regressive overlap data, seismic sequence stratigraphy (Haq, 1991) and the use of glacio-hydro-isostatic rebound models. In reality no single measurement is definitive and a characteristic of these lines of investigation is their frequent interdependence. Thus, whilst each approach provides useful calibration data, their interdependence makes truly independent testing of one or other measurement difficult.

Ice sheet reconstructions and ocean volume

The link between ice sheet volume (specifically terrestrial ice) and ocean volume is an obvious one, and ice sheet reconstructions from the LGM to the present provide a measure of the contribution of meltwater to ocean volume change during this interval. The most widely known example of this approach is that by Denton and Hughes (1981), who presented maximum and minimum ice model estimates of Late Quaternary ice sheet reconstructions (Table 1.1).

Initially designed to provide boundary conditions for the CLIMAP project, almost every estimate has been the focus of intense scrutiny and revision since their publication. One of the main limitations of the ice sheet reconstructions of Denton and Hughes (1981) and others is their heavy dependence on reconstructions of former ice sheet surface profiles. In some cases these reconstructions are constrained by trim lines which mark the altitudes at which nunataks protruded above the ice surfaces (cf. Ballantyne *et al.*, 1997). However, in the case of the largest ice sheets which developed over relatively flat terrain, such as the Laurentide, nunatak evidence is often lacking. In these cases, the surface profiles can only be estimated by mathematical models of ice flow, which introduce considerable uncertainty in the estimates of ice sheet volume.

Oxygen isotope records and ocean volume

Oxygen isotope data of benthic and planktonic foraminifera reflect, amongst other factors, large-scale changes in rates of continental ice sheet growth and decay. Shackleton (1987) argued that a 0.1‰ change in ^{18}O is approximately equivalent to a 10m change in global sea level and on this basis produced a sea level curve for the last 160,000 years (Figure 1.9, Chappell and Shackleton, 1986), although the data more properly represents a curve of changing ocean volume over time. Complications in this approach include temperature and salinity-dependent changes in isotopic composition of the foraminifera, the ubiquitous degradation of deep sea cores by bioturbation, as well as changes in composition of ice sheets as they expand and contract, and related changes in the isotopic composition of seawater over time.

TABLE 1.1 Relationship between volumes of last ice sheets and global sea level lowering according to the minimum and maximum ice reconstructions of Denton and Hughes (1981). The values assume isostatic equilibrium and a rock-ice ratio of 4

	Minimum Reconstruction		*Maximum Reconstruction*	
Ice Volume	*Total Volume (10^6 km^3)*	*Volume Causing Lower Sea Level (10^6 km^3)*	*Total Volume (10^6 km^3)*	*Volume Causing Lower Sea Level (10^6 km^3)*
Ice Sheets				
Laurentide	30.900	30.500	34.800	34.200
Cordilleran	0.260	0.260	1.900	1.840
Innuitian			1.130	0.983
Greenland[a,b]	2.920[a]	0.287[b]	5.590[a]	2.550[b]
Greenland[b,c]	0.287[c]	0.287[b]	2.950[c]	2.550[b]
Iceland	0.050	0.050	0.267	0.236
British Isles	0.801	0.773	0.801	0.773
Scandinavian	7.250	7.060	7.520	7.320
Barents-Kara	0.955	0.865	6.790	6.250
Putorana			0.581	0.581
Antarctica	37.700	9.810	37.700	9.810
East	24.200	3.330	24.200	3.330
West	13.500	6.480	13.500	6.480
Glaciers and ice caps	1.184	1.830	0.750	0.715
Totals	84.174	51.300	97.829	65.300
Sea level equivalent[d]		127m		163m
Eustatic sea level drop[e]		91m		117m

After Denton and Hughes (1981)

[a] Total Late Wisconsin-Weichselian ice.

[b] Additional Late Wisconsin-Weichselian ice contributing to lower Late Wisconsin -Weichselian sea level.

[c] Additional Late Wisconsin-Weichselian ice.

[d] Without hydro-isostatic sea floor rise, using present ocean area (361×10^6 km^2).

[e] With hydro-isostatic sea floor rise.

Initially, benthic foraminifera analysed by Shackleton (1977) suggested that sea level at the LGM fell by as much as 160m (which equates to 55-60 x 10^6 km^3 of water) in line with the maximum estimates of equivalent sea level change proposed by Denton and Hughes (1981). However, Shackleton (1987) subsequently refined this curve by using a combination of benthic and planktonic foraminifera to correct for temperature-dependent changes in ^{18}O composition, and concluded that the sea level minimum at the LGM was less than previously thought (*circa* -120m).

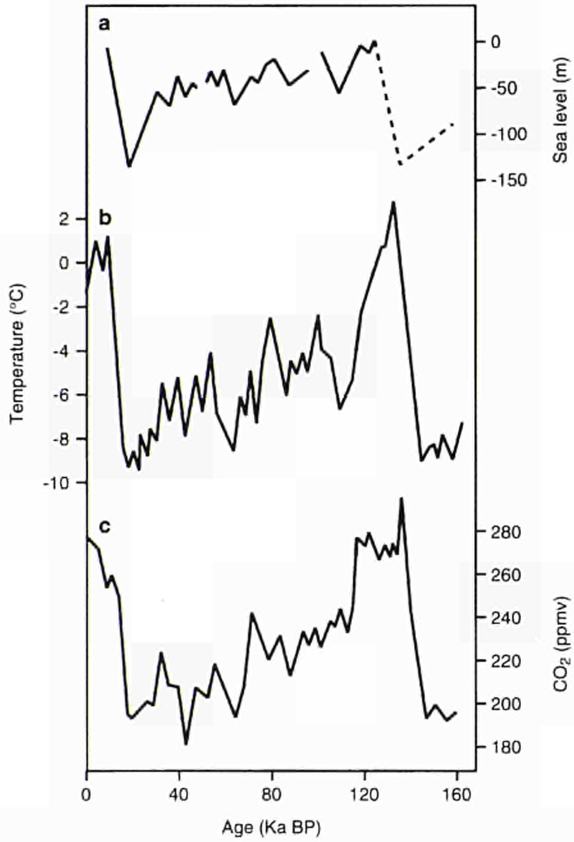


FIGURE 1.9 Sea level changes (a), variations in temperature (b) and changes in CO_2 for the past 160ka (Chappell and Shackleton 1986, Jouzel et al., 1987 and Barnola et al., 1987, redrawn in Tooley 1993).

This revised curve has a close resemblance to sea level curves produced from the raised coral reefs of Papua New Guinea (Figure 1.10). Of particular significance is the demonstration that ocean volume and sea level during the last interglacial (isotope substage 5e) was greater than at any time during the present interglacial, the latter by between 1 and 5m.

Shoreline data and sea level change

Shoreline data, such as those used in the reconstruction of the Papua New Guinea sea level curve, are widely used in the reconstruction of sea level change although their interpretation is almost universally complicated by problems of dating (see below) as well as difficulties involved in disentangling the relative contribution of glacio-isostasy, hydro-isostasy, tectono-eustasy, geoidal-eustasy as well as vertical crustal movements.

Few records of sea level change since the LGM exist and those that do (e.g. from Papua New Guinea, Barbados and Tahiti) rely on both emerged and submerged corals in regions remote from the locations of former ice sheets. Data from these regions are interpreted by some as

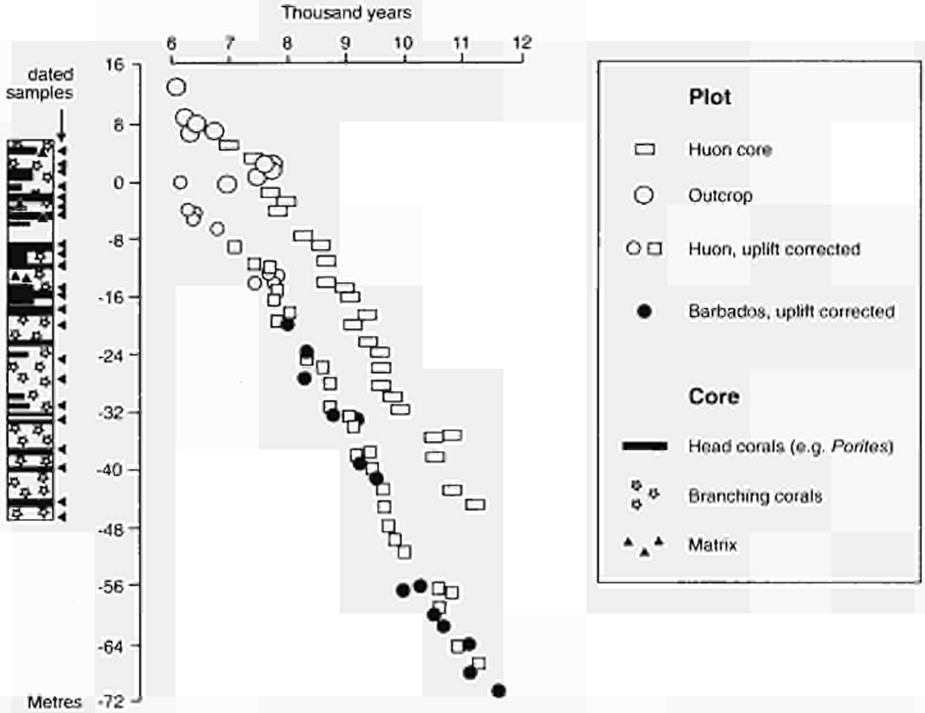


FIGURE 1.10 Comparison of the age and altitude of radiocarbon-dated curves from the Huon Peninsula, Papua New Guinea and Barbados, corrected for uplift (Chappell and Polach 1991).

reliable records of globally averaged sea level. However, this is unlikely to be the case for a number of reasons:

a) It has been proposed by Peltier (1994) that during deglaciation, water was transferred poleward from equatorial regions in order to ensure that the geoid remained coincident with a gravitational equipotential. A consequence of this interpretation is that there was a reduced net rise in sea level at the Huon Peninsula (Papua New Guinea) compared to Barbados of approximately 15m. This led Peltier (1994) to conclude that the Barbados sea level curve would be an over-estimate if used as a globally averaged measure of sea level rise.

b) It is possible that both Barbados and the Huon Peninsula have experienced episodic tectonic movements. However, the sea level data from these areas are corrected for an average uplift rate of 0.34mm yr^{-1} (Barbados) and 1.9mm yr^{-1} (Huon Peninsula). Only the mid-Pacific island of Tahiti (Bard *et al.*, 1996; Montaggioni *et al.*, 1997) is far from any plate boundary.

c) All three areas lie within 'intermediate field' locations (Lambeck, 1993a), meaning that although distant from former ice sheets, nevertheless their sea level histories will record additional glacio- and hydro-isostatic effects related to ice sheet melting and the redistribution of water.

d) Finally, much of the Caribbean records are based on a species of *Acropora palmata*, a coral that generally forms in waters less than 5m depth. This error associated with the water depth in which *Acropora palmata* presently exists implies thus the reconstructed pattern of relative sea level changes produced both by Fairbanks (1989) and Blanchon and Shaw (1995) for the time period from *circa* 17,000 years BP till the present day include this source of error.

1.2.2.5 Rates of ocean volume and sea level change. Despite the above limitations, the coral reef sea level records provide important, though highly debated, insights into the rate of ocean volume change since the LGM. For example, the Barbados record shows sea level rising from a minimum of -121$\pm$5m with two pronounced steps at *circa* 14k cal years BP and 11.3k cal years BP (Figure 1.11).

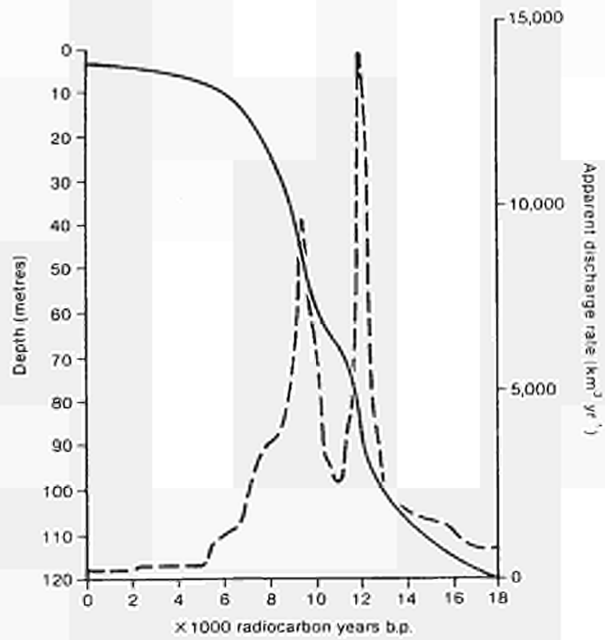


FIGURE 1.11 The Barbados sea level curve (Solid line) based on an analysis of dates from submerged corals and overlaid by a dashed line showing apparent meltwater discharge rates, the peaks of which coincide with periods of enhanced sea level rise (Fairbanks 1989, redrawn in Tooley 1993).

Fairbanks (1989) linked these accelerations in sea level rise to meltwater pulses into the Gulf of Mexico and the North Atlantic. More recently, Blanchon and Shaw (1995) used these and other coral reef data from the Caribbean-Atlantic region to identify three catastrophic sea level rise events (termed CRE-I to -III) (Table 1.2). These rises in sea level were, it was argued, synchronous with the collapse of the Laurentide and part of the West Antarctic ice sheets, together with major reorganisation of ocean-atmosphere circulation, and large-scale releases of subglacial and proglacial meltwater.

TABLE 1.2 Catastrophic sea level rise events from the Caribbean-Atlantic region (after Blanchon and Shaw, 1995)

Event	Start (Cal Years BP)	Magnitude (m)	Duration (cal years)
CRE-I	14.2±0.1	13.5±2.5	<290±50
CRE-II	11.5±0.1	7.5±2.5	<160±50
CRE-III	7.6±0.1	6.5±2.5	<140±50

Although important aspects of the Blanchon and Shaw (1995) chronology have been questioned (Clark, 1995, see below), significant reservations have also been raised in a recent coral sea level study from Tahiti by Bard *et al.* (1996) and Montaggioni *et al.* (1997). It has been known for some time (e.g. Broecker, 1993) that the Barbados curve of Fairbanks (1989) was based on samples from three cores, within each of which the rate of sea level rise is nearly constant. However, the record from each core is offset, and it is these offsets which form the two apparently rapid meltwater surges identified at *circa* 14k cal years BP and 11.3k cal years BP. Bard *et al.* (1996) observe a similar problem with the more recent event CRE-II (7.6k cal years BP) identified by Blanchon and Shaw (1995), which coincides with the switch in data sets from Barbados drill sites to other samples collected from the Caribbean by divers. More importantly, the record described by Bard *et al.* (1996) is based on two overlapping cores (for the last 10ka cal years BP) and this enables appraisal of between core variations in reef growth characteristics. Comparison of the two records led Bard *et al.* (1996) to suggest that local reef responses (i.e. recorded in one but not both cores) are significant, and that the growth history they record is one controlled by the progressive flooding of the Tahiti islands. These authors find indirect support for the first meltwater phase identified by Fairbanks (1989). However, they find little evidence for the two younger reef drowning events CRE-II and CRE-III of Blanchon and Shaw (1995) which, if real, "were only regional occurrences and, consequently, cannot have been triggered by global glacio-eustatic perturbations" (Montaggioni *et al.*, 1997, pp557).

Dating problems also beset CRE-III, which is presumed to equate with the catastrophic disintegration of the Laurentide ice, the creation of a floating "ice-Island" over Hudson Bay and the discharge of glacial meltwater into the Labrador Sea via The Hudson Strait. The timing of this catastrophic Laurentide ice sheet collapse is presently estimated from the age of marine shells deposited in the Tyrrell Sea as the sea invaded deglaciated areas of the proto-Hudson Bay. The shell dates (reservoir corrected) provide ages of *circa* 7,600 ^{14}C years BP (Dyke, pers. comm.) and therefore correspond well with the age of the CRE-III sea level rise calculated by Blanchon and Shaw (1995) (Vincent and Hardy, 1979). It would therefore appear, according to the Blanchon and Shaw (1995) data, that the final disintegration of the Laurentide ice sheet was indeed accompanied by a catastrophic sea level rise event. There is, however, evidence from the continental shelves and at present on land sites of periods of accelerated sea level change including sea level rise during the last 100ka. It has been suggested that periods of accelerated sea level rise lasted up to 450 years and rates ranged from 10-75mm yr^{-1} (Cronin, 1983; Tooley, 1978, 1989, 1994; Tooley and Jelgersma, 1995).

1.2.2.6 Glacio-hydro-isostatic modelling and ocean volume change. Although geophysical rebound models rely in part on estimates of ocean volume change, they are also capable of testing previously proposed models of ice sheet reconstructions by comparing modelled and observed relative sea level changes. The ice sheet reconstructions pose particular problems since they reflect two different theories on the ways that ice sheets develop. The Denton and

(1981) hypothesis is that the last ice sheets of the northern hemisphere (specifically the Laurentide and Fennoscandian ice sheets) were associated with single domes. The rationale of this argument is based on the view that because the oxygen isotope records indicate a global fall in sea level of more than 75m during the last interglacial-glacial transition (i.e. between *circa* 124,000 and 115,000 years BP) and because most glaciological models cannot generate the necessary ice volume in the time required, extensive ice shelves must have developed during this period in Hudson Bay, the Gulf of Bothnia, the Canadian Arctic archipelago and the Arctic Ocean. Thereafter, these ice shelves may have become grounded and thus enabled the growth of large ice sheets. By contrast, many scientists (e.g. Andrews, 1982) criticised this "outrageous hypothesis" and pointed out that the field data indicate quite different styles of ice sheet growth characterised by initial ice accumulation on upland plateaux and the eventual development of multi-domed ice sheets that were significantly smaller than those envisaged by Denton and Hughes (1981). Of significance here is the fact that these two radically different ice sheet models are associated with large differences in the estimates of ice volume for the LGM.

Geophysical rebound models can test these contrasting ice sheet reconstructions. Two examples of this approach are given here for the Barents-Kara Sea area, and the British and Fennoscandian ice sheets.

The Denton and Hughes (1981) and Grosswald (1980) maximum ice sheet reconstruction envisaged a large ice sheet positioned over the Barents-Kara Seas which would have coalesced with the Fennoscandian ice sheet and have been broadly equivalent in size. An alternative and more widely held view is this ice sheet was much smaller, restricted to the Svalbard archipelago and the Russian Arctic coastal zone (Boulton *et al.*, 1985; Velichko, 1984; Tveranger *et al.*, 1995; Mangerud *et al.*, 1996). Recent field evidence has leant support to the presence of a large ice sheet over the Barents Sea and restricted ice cover over Arctic Russia (Solheim *et al.*, 1990; Elverhoi *et al.*, 1993) although uncertainty remains concerning the exact dimensions of this ice sheet and whether it comprised one or more domes. Using maximum and minimum ice sheet models, Lambeck (1995a) finds strong support from the empirical sea level data for a substantial ice sheet over the Barents Sea with a maximum ice thickness of 1500-2000m. Although the ice may have extended over the Kara Sea, it must have been limited in areal cover and substantially thinner than presumed in the maximum reconstructions. Forman *et al.*, (1996), in a discussion of the sea level data on which Lambeck's (1995a) computations depend, suggest that their new data indicate a reduced global sea level contribution from ice sheets in the Russian Arctic continental shelves to between 6 and 10m, which is significantly less than the previous estimates of 16m (e.g. Nakada and Lambeck, 1988).

Another example of the way in which rebound models are used to refine ice sheet size and hence potential ocean volume contribution is provided by recent debate concerning ice thickness at the time of the UK LGM. These vary by 25-50% with some models showing the UK and Fennoscandian ice sheets merging whilst others do not (e.g. Boulton *et al.*, 1977, 1985; Denton and Hughes, 1981; Sejrup *et al.*, 1994). Using maximum and minimum models, Lambeck (1991, 1993b) has shown that sea level observations from Scotland are inconsistent with a coalescing ice sheet model, unless this occurred prior to *circa* 23k ^{14}C years BP, which he regarded as much earlier than previously considered by other authors.

These two examples show how rapidly developing ideas about the size and chronology for ice sheet growth and melting are altering our understanding of the relative contribution of different ice sources to changes in ocean volume since the LGM. It should also be noted that very little information is available for the ice histories of Antarctica and Greenland since the last LGM and, given the possible uncertainties of the ice reconstructions for these two areas, much of the discussion concerning ice sheets as small as that which covered the UK may be largely academic. The ICE-4G model which Peltier (1994) proposes suggests a global LGM ice sheet palaeotopography very different to that advocated by Denton and Hughes (1981). Indeed, Peltier (1994) suggests that both the maximum and minimum estimates of Denton and Hughes (1981) are too large, the former by *circa* 55%. Significant differences in climate reconstructions using global atmospheric circulation models are therefore anticipated between the original CLIMAP predictions and future estimates based on the Peltier (1994) model.

1.2.2.7 Discussion. The above review highlights areas which warrant further research. The first revolves around the cross-cutting theme of the dating, which is essential for establishing the chronology of ice sheet dynamics and sea level change. The second relates to the geographical and temporal deficiencies in the available information regarding ice sheet histories and patterns of relative sea level change since the LGM. Improving knowledge is fundamental to the marine oxygen isotope studies as well as rebound models and chronologies of ice sheet decay.

Problems in the Reconstruction of Chronology

It is well-known that the calibration of the radiocarbon timescales for the Holocene is linked to the tree-ring record of climatic variations (Kromer *et al.*, 1995). Recent attempts to extend the tree-ring timescale have succeeded in producing an absolute dated tree-ring calendar back to 11,597 BP (calendar years before AD1950) (Hajdas, 1993; Bjorck *et al.*, 1996). However, for the Lateglacial period the uncertainties in establishing a chronology are considerably greater. In the absence of a tree-ring timescale, a varve chronology and a partially established Uranium/Thorium chronology linked to AMS dates of laminated lake sediments is used (Wohlfarth *et al.*, 1995). The Uranium/Thorium data of corals, cross-matched with radiocarbon ages, provides only a very approximate relationship between calendar years and radiocarbon years for the Lateglacial (Bard *et al.*, 1993). This is a particularly crucial factor in understanding Lateglacial ocean volume changes, since it is during this time that the greatest changes in relative sea level took place. This has an added significance since patterns and rates of Holocene relative sea level changes for different areas of the world very much depend upon the nature and style of the flux rates of glacial meltwater into the world's oceans during the Lateglacial (see above).

In part this problem is being addressed through studies of Greenland ice cores (Gronvold *et al.*, 1995) where volcanic tephra layers dated to a specific calendar year can be identified not only in ocean floor sediments but also within Lateglacial varve sequences. This has a particular importance in respect of the (Icelandic) Vedde ash layer produced by a large eruption c. 11,900±80 cal. yrs BP that can be traced in various North Atlantic ocean floor sediments, bit also within Swedish varve sequences. The tracing of this tephra horizon together with younger volcanic tephra layers may be used to establish a time-synchronous stratigraphic horizon across NW Europe that, in turn, may be used to check the calendar age chronology for the Lateglacial. This research, is additionally complicated by changes in the reservoir age of ocean carbon related to N Atlantic deep water formation. For example, Austin *et al.* (1994) has demonstrated that whereas the average apparent age of modern mollusca in the N Atlantic region is 440 years, it may have been the case that the apparent age of seawater during part of the Lateglacial was in the order of c. 800 years. At present, possible variations in the apparent age of seawater during the Lateglacial are not known. This is particularly important in respect of Lateglacial sea level reconstructions that make use of dated marine mollusca in order to reconstruct former sea level and ice marginal positions.

Dating problems also restrict our understanding of the most complete records of local relative sea level change that have been produced for the Lateglacial and Holocene. As noted above, Clark (1995) has commented on problems associated with the chronology established by Blanchon and Shaw (1995) and drew attention to particular difficulties in linking their radiometric age estimates of CRE events to the calendar year timescale. More recently, Bard *et al.* (1996) noted that the slowdown in relative sea level rise during the Younger Dryas recorded for the Caribbean is not registered in the Tahiti coral record. it is within the perspective of these age estimate errors and the considerable difficulties in linking the radiocarbon and calendar year timescales for the Lateglacial, that our reconstructions of former changes in ocean volume flux are determined.

Additional complications are posed in respect of earth rheological models where the model predictions of Lambeck rely on published radiocarbon ages of variable quality. The Peltier models are similarly dependent at present on the relatively crude calibrations that exist between the Uranium/Thorium timescale and the calendar year timescale. These sources of error, in turn, introduce themselves into the result of the rheological models used to reconstruct trends in sea level.

1.2.3 Global sea level variations from tide gauge observations

Philip Woodworth

Coastal tide gauges have provided the main technique by which sea levels have been measured during the past century (for description of their operation, see Chapter 2, section 2.2.2). Tide gauge data from many countries around the world have been collected routinely by the Permanent Service for Mean Sea Level (PSMSL), which is a member of the Federation of Astronomical and Geophysical Data Analysis Services (FAGS) and which operates under the auspices of the International Council of Scientific Unions (ICSU). The data bank holds approximately 43000 station-years of monthly and annual values of Mean Sea Level (MSL) from over 1750 stations worldwide. Where possible, records at each site are placed into a Revised Local Reference (RLR) data set, wherein MSL values at a station are referred to the same reference height (i.e. the 'RLR' datum which is defined in terms of the height of the tide gauge benchmark or TGBM). Only RLR records can be used for time series analysis, although all MSL stations-years (called 'Metric' data in PSMSL terminology) can be used for studies of seasonal cycles.

Inspecting the geographical distribution of PSMSL data (Woodworth, 1991), it appears at first sight that copious amounts of information are available from virtually every point on the world coastline (Figure 1.12).

However, a closer inspection shows that many records are quite short. A requirement that records be more than 20 years long loses most stations in Africa and many at ocean islands. Owing to the existence of significant interannual and interdecadal variability in many records, Douglas (1991), for example, stipulated the requirement of record lengths of 60 years or more for the calculation of a reliable long term trend. This results in stations for only northern Europe, North America and Japan, along with odd ones in the southern hemisphere such as Sydney or Buenos Aires. Therefore, it is important to keep in mind that the 'global' sea level data set from the past century is not only just a coastal set, but is also

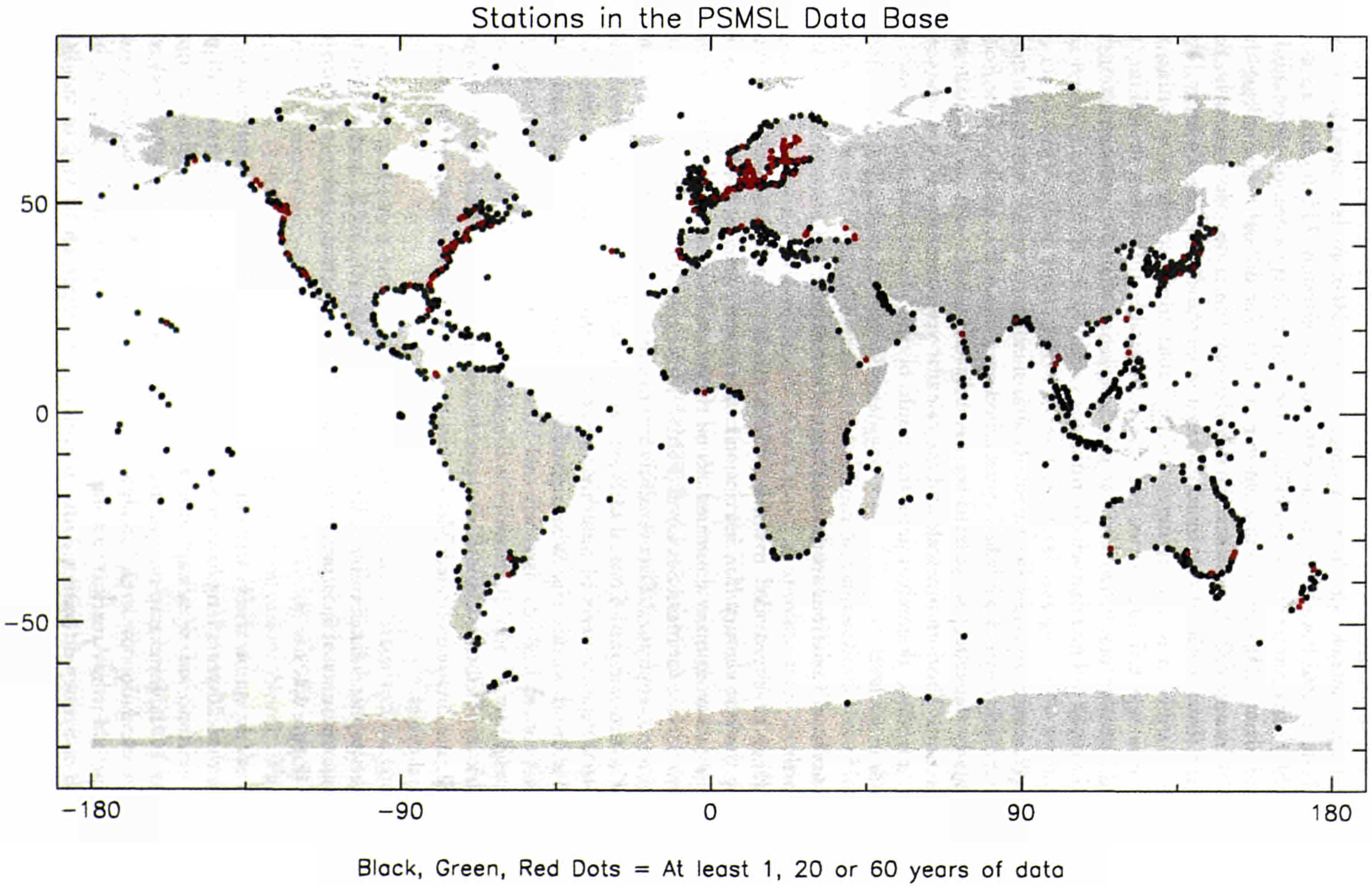


FIGURE 1.12. The distribution of tide gauge stations in the PMSL data set, with at least 1 (black), 20 (blue) or 60 (red) years of data.

primarily just a northern hemisphere one. There is also spatial variability in the quality of the records; Woodworth (1997) reviews aspects of the quality of the historical data set including the accuracy of long term trends, the need for good data quality control (e.g. via 'buddy checking') and for good local geodetic control to guard against the possibility of unexpected very local land movements (e.g. submergence of the gauge itself, perhaps on the end of a pier) propagating into the long term record.

The spacio-temporal distribution and varying quality of the tide gauge records leads to problems in the estimation of global mean sea level changes over the last century; approaches to these problems are discussed in more detail in the following paragraphs. Recent researchers of the long tide gauge records in the PSMSL data set have obtained estimates for the twentieth-century trend in global sea level of approximately 18 cm/century (+/- 7 cm/century). For reviews, see the Second Scientific Assessment of the Intergovernmental Panel on Climate Change (IPCC) (Warrick *et al.*, 1996) and Douglas (1995). This is, perhaps, a reassuring result, although it has to be kept in mind that all authors have used the same (PSMSL) data source.

The main problem analysts have to face, aside from the geographical inhomogeneity of the data set, is the need to remove vertical land movement signals from the gauge records, thereby converting time series of 'relative' sea level change (relative to the height of the land on which the gauge is situated) to 'real' sea levels. Different authors have approached this problem in different ways. Peltier and Tushingham (1989, 1991), Trupin and Wahr (1990) and Douglas (1991, 1997) used versions of Peltier's geodynamic models of post-glacial rebound (PGR). Of course, PGR is not the only geological contribution to vertical land movements, but it is the only one for which we possess detailed understanding, i.e. for which we have a model capable of being employed on a global basis (Peltier and Tushingham, 1989; Lambeck, 1990). However, the subject of parameter values to be used within PGR models is a topic currently being discussed intensively (Mitrovica and Davis, 1995; Peltier, 1996). Douglas, in particular, was careful to reject tide gauge records from stations which he considered to be outside of the areas for which PGR is the dominant geological process, and at which, therefore, he could not make a reasonable attempt to estimate the vertical movements.

Gornitz and Lebedeff (1987) and the European regional analysis of Shennan and Woodworth (1992) took a different approach, using directly in their analyses those sets of geological information of different ages obtained from around the gauge sites, in order to extrapolate the Holocene sea level curves into the present day when they can be considered as primarily reflecting very long timescale geological change. This procedure, in principle, extrapolates all the vertical land movement signal (other than, of course, rapid changes such as due to earthquakes), whether mostly PGR or not. However, it appears to result in systematically lower values for the estimated twentieth-century sea level trend; for a fuller discussion, see Warrick *et al.* (1996).

Whatever the details of the analysis, it is clear that most long tide gauge records from around the world show evidence for increasing levels (Woodworth, 1991). It is interesting, however, that some of the longest, and highest quality, records are from northern Europe (e.g. Stockholm, the longest continuous record in the world, Ekman (1988)). These have not so far been employed by most analysts in global studies as the 'near field' accuracy of the PGR models has not been adequate to perform a meaningful subtraction from the tide gauge records, unlike the 'far field' situation exploited by Douglas (1995) and others.

Towards improving the data base, in future, the vertical land movements at gauge sites will be monitored by means of new geodetic techniques, primarily the Global Positioning System (GPS) and absolute gravity (Carter *et al.*, 1989; Carter, 1994; Neilan and Woodworth, 1997). The tide gauge and GPS communities have proposed a 'medium term strategy' whereby geodetic measurements might make maximum benefit of the historical tide gauge data set, by prioritising measurements at sites with long gauge records (IOC, 1997; Neilan and Woodworth, 1997). At sites which prove to have 'linear geological trends' (i.e. linearly, or at least monotonically, changing vertical land movement, with no large excursions owing to earthquakes etc.), with a standard error on the GPS-derived vertical land movement trend less than that of the tide gauge trend, the GPS trends can be used to hindcast the vertical land movements within the historical records. (In practice, this might mean a GPS record of land movements approximately 10-20 years long being applied to a tide gauge record 60 years long, see Woodworth, 1997).

At those sites where it is clear that the geological trends are not 'linear' (e.g. some eastern Mediterranean sites), recording has in effect to start again with GPS measurements taken in parallel to the tide gauge data. The benefits of such investment in terms of obtaining more spatially-representative trends will clearly take longer to be realised, although with geophysical insight it is feasible that studies may provide acceptable limits to real sea level trends over reasonable periods.

If GPS, and the other new geodetic techniques, had not been invented, tide gauge analysts would obviously have continued to study sea level variations. The subject would still have developed through improvements in geodynamic models and their application to studies of trends, and through monitoring of any 'accelerations' at sites with the longest records. Various indices can be computed which attempt to represent 'accelerations' (or anomalous departures from predicted levels) using the assumption that geological change at many sites is essentially linear with time. For example, Shennan and Woodworth (1992) present a 'sea level index' for the North Sea area indicating an apparent recent fall in levels over decadal timescales.

Long term 'accelerations' in sea level trends are of interest from several points of view. First, the 10-25 cm/century average trend estimated for the past century is apparently larger than one would expect based on the average trend over the past two millennia (see references in Warrick *et al.*, 1996), although previous century-timescale variability could certainly have occurred within that longer period. For example, there is evidence that sea level rise in the Mediterranean during Roman times was greater than that of the more recent past (Pirazzoli, 1976). However, accelerations measured over the past two centuries using the longest European tide gauge records (Woodworth, 1990) appear to be small, so it is unclear when the current trend might have commenced. Certainly, there is no evidence for significant accelerations over century or many-decade timescales using data from this century alone (Woodworth, 1990; Douglas, 1992), which puts into context the potential importance of a several-fold increase in trend during the next century (Warrick *et al.*, 1996), especially if that acceleration is greater in the North Atlantic (Mikolajewicz *et al.*, 1990).

One incentive for making reliable observation based estimates of global mean sea level change over the past century is to provide a check on model based estimates of changes in the volume of the ocean resulting from climate forcing. A consistency between observation and model based estimates for the past would lead to greater confidence in our future predictions of climate change related sea level change. In other words, we need to know if our representation of the physics of the climate system is essentially correct within the models used to estimate sea level changes.

An example of the 'need to know more' is given by the fact that we do not expect long term sea level changes (whether 'trends' or 'accelerations') to be the same everywhere because of changes in the ocean circulation and ocean loading. In Table 11 of Douglas (1991), one sees a remarkable uniformity in trends observed at most locations over the past century (although the uncertainties could also accommodate a difference of a factor of two between the trends of the European Atlantic and eastern North American coasts). However, this apparent uniformity need not be the case with regard to future changes. GCM based estimates of future sea level changes due to thermal expansion show large spatial variation. For example, in Gregory (1993) the eastern coast of North America shows larger than average rise while the area to the north of the Ross Sea shows constant sea level or even a fall, such features are also indicated in GCM runs by other authors. Conrad and Hager (1997) show that spatial variations due to ocean loading changes caused by the present-day melting of glaciers and ice sheets lead to a 5-10% uncertainty in tide gauge based estimates. Other possible spatial variations in future trends have been suggested by results from the reduced gravity model studies of Hsieh and Bryan (1996), which indicate that advanced signals of sea level rise can propagate rapidly through the action of Kelvin and Rossby waves, but that the full adjustment toward a more uniform sea level rise may take place much more slowly. The models indicate how any rise owing to ocean heating could be difficult to estimate from coastal stations alone. The 'climate fingerprint' of the GCM's and the deep ocean/coastal differences of Hsieh and Bryan can only be identified if we have the fullest coverage of sea level trend measurements, both by coastal tide gauge and by altimetry in the deep ocean. This 'climate fingerprint', if real, can only be isolated if we can measure the real sea level trends, both by coastal tide gauges and by altimetry in the deep ocean.

In the next decade, programmes such as the Global Sea Level Observing System (GLOSS) (IOC, 1997), and its regional projects (e.g. Baker *et al.*, 1997), should enable a continued improvement in the geographical coverage of long tide gauge records, and in the quality of those records. Many tide gauges will become combined gauge/GPS/met stations, capable of providing much more of the information which oceanographers and geodesists require. Gauges will continue to play major roles in monitoring aspects of the changing ocean circulation, and will provide an on-going calibration capability to a series of altimeter missions (Mitchum, 1997; IOC, 1997).

1.2.4 Global sea level variations from satellite altimetry

Anny Cazenave

1.2.4.1 A global mean sea level curve. Topex-Poseidon is the first altimetry mission specifically designed and conducted for studying the large scale ocean circulation (Fu *et al.*, 1994, 1996; see also the collection of papers in two special *Journal of Geophysical Research* issues, vol. 99, Dec. 1994 and vol 100, Dec. 1995). (For a schematic diagram illustrating how the sea surface is measured using satellite altimetry, see Chapter 2, section 2.2.4, Figure 2.1).

The accuracy presently reached by Topex-Poseidon in measuring the sea surface height, has allowed the time evolution of the sea level to be monitored within 1 mm/yr. Using the first 2 years of Topex-Poseidon data, several investigators have attempted to determine the rate of mean sea level change (Nerem, 1995a, 1995b, Minster *et al.*, 1995, Hendricks *et al.*, 1996).

Unfortunately in these early investigations which reported a large global mean sea level rise of 5-7 mm/yr, the data were contaminated by a drift of the on-board Topex oscillator, incorrectly taken into account in the correction algorithm.

Most recent results based on corrected data (e.g., Nerem *et al.*, 1997a, 1997b, Cazenave *et al.*, submitted, P.Y. Le Traon, personal communication) suggest a significantly reduced mean sea level rise, <1.5 mm/yr. Using recently reprocessed, high-quality Topex-Poseidon data, P.Y. Le Traon (personal communication) and Cazenave *et al.* (1997) report a global mean sea level rise of ~1.2 $\pm$ 0.15 mm/yr for the period January 1993 through July 1997. From the first 4 years of the Topex-Poseidon mission, Nerem and co-workers found initially rates in the range −0.2 mm/yr to +0.5 mm/yr (Nerem *et al.*, 1997a, 1997b). However their most recent global mean sea level slope now agrees with that of Le Traon and Cazenave *et al.* (S. Nerem, personal communication, December 1997).

The global mean sea level curve (after Cazenave *et al.*, 1997) derived from averaging individual sea surface height measurements (equi-area weighted) between $\pm$ 60° latitude over each Topex-Poseidon cycle (~10 days) is shown in Figure 1.13. Since changes in the ocean circulation should average to zero globally because of mass conservation, this mean sea level curve should essentially represent ocean mass and volume changes, i.e., exchange of water mass with continents (including ice sheets) and steric effects. As noted earlier by Nerem (1995a, 1995b), there is some correlation between sea surface temperature (SST) and sea level changes from monthly to interannual time scales. Such a comparison is presented in Figure 1.13 where the SST curve (semi-annual and annual components removed) is superimposed on the sea level curve. It is interesting to note that both sea level and SST time series present step-like discontinuities in autumn 1994 and early spring 1997, likely related to the 1994 and 1997 ENSO events. The correlation between sea level and SST is interesting and further suggests that the observed global sea level changes from intra seasonal to interannual time scales result mainly from steric effects, in relation to variations of heat storage in the surface ocean.

Mitchum (1997) compared sea surface height time series from Topex-Poseidon with sea level data from tide gauges. A subset of the global tide gauge network was selected for this purpose, mostly located within the $\pm$ 30° latitude band. Over the years 1993-1996, Mitchum reports a drift of -2.3 mm/yr between the altimeter and the tide gauge-derived mean sea level, further attributed to some still unknown drift of the altimeter measurements. Although this trend of −2.3 mm/yr cannot be simply subtracted from the Topex-Poseidon-derived sea level rate for a number of reasons (the tide gauges favor the tropics, they are rarely located on the satellite ground track, tide gauge data are not corrected for vertical land movements, etc), it points out that some remaining instrumental drift affecting the altimeter data cannot be excluded (e.g., a possible drift at the mm/yr level of the on-board microwave radiometer used to estimate the wet tropospheric delay, is currently investigated).

1.2.4.2 What about the future? Satellite altimetry records are of course still too short to detect long term sea level rise caused by possible global warming, and the results discussed above clearly do not compare to those based on multi-decadal tide gauge records. However satellite altimetry clearly has an advantage over the latter, i.e., to routinely monitor the 'absolute' sea level with a nearly global coverage, and hence to be able to record as time progresses, the 'true' volume and mass changes of the oceans. Measuring global sea level changes of climatic origin is probably the most challenging application of satellite altimetry. Crucial for this objective, is the need for future altimeter satellites of the Topex-Poseidon-class in order to acquire long and accurate time series of global sea level. The recently decided JASON-1 mission makes hopefully this objective achievable.

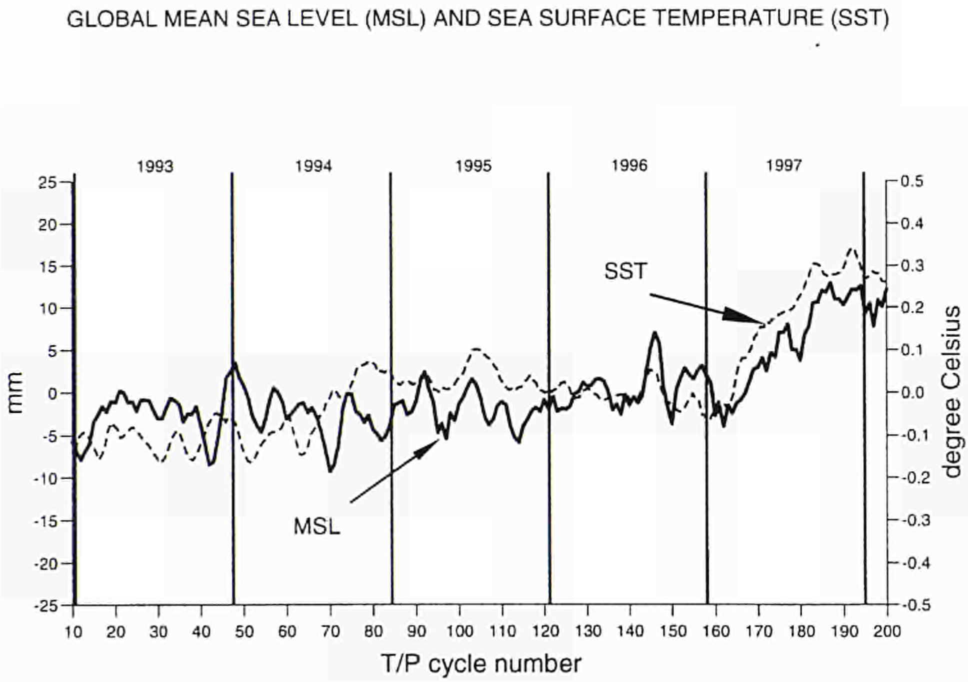


FIGURE 1.13 Global mean sea level and sea surface temperature (after Cazenave et al., 1997)

1.2.4.3 Regional variations. Since the sea level height measurements are provided every few km along the satellite tracks and since the ground track coverage is homogeneous over the globe (the inter-track spacing is ~350 km at equator for Topex-Poseidon), it is possible to map the regional pattern of the sea level slopes estimated over the few years of altimetry data available.

The rate of sea level change (for 1993-1996) as a function of geographical location (Figure 1.14) shows considerable regional variations ranging from ~+ 50 mm/yr to ~-50 mm/yr. It is interesting to notice that while in some areas sea level is rising (e.g., in the Western Equatorial pacific, North Pacific, South Atlantic), in other regions, sea level is dropping (e.g., Eastern Pacific). The spatial pattern shown on this map is dominated by the regional inter-annual variability of the ocean circulation. Moreover, the positive sea level anomaly seen in the Western Pacific is likely related to the prolonged series of ENSO events which occurred since the early 1990s, in particular in 1992 and 1994. On the other hand, the geographical pattern of the SST drift (not shown) shows some striking correlation with that of the sea level, in particular in the Pacific, an indication that part of the regional sea level changes are related to heating or cooling of upper ocean layers. The pattern shown in Figure 1.14 confirms that sea level variations associated with temperature changes are not at all uniform, neither in the open ocean nor along coastal regions.

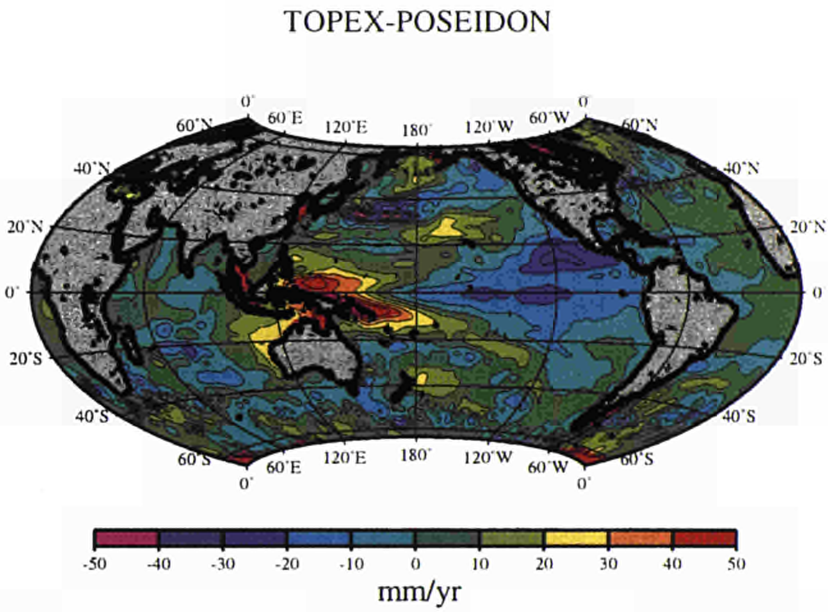


FIGURE 1.14 Geographical variation of sea level rise from Topex-Poseidon (1993-1996)

1.3 ESTIMATING THE CONTRIBUTIONS TO OCEAN VOLUME CHANGE FOR THE LAST AND NEXT 100 YEARS: OBSERVATIONS AND MODELLING

1.3.1 Introduction

To comprehensively assess changes in the volume and mass of the ocean it is necessary to consider the whole hydrological cycle though, to date, thermal expansion and the melting of land-based ice are generally thought to be the most important contributors for the next century. These are the factors upon which we concentrate here because studies of the past contribution of other factors are few and they have not been considered in most future projections. However, we acknowledge the need for further research in these areas. Sahagian *et al.* (1994a) estimate the present direct anthropogenic contribution from a combination of groundwater withdrawal, surface water diversion and land-used changes to be 0.54 mm a^{-1}, which they regard as a minimum estimate. But this paper leads to considerable debate in particular about the amount of water storage as a result of dam building and the plans for dam building in the future (Greuell, 1994, Sahagian *et al.*, 1994a, b & c, Chao, 1994, Rodenburg, 1994). A recent review by Gornitz *et al.* (1997) preliminarily estimates that the net effect of human intervention in the hydrological cycle is currently -0.8±0.4 mm/y. In addition, climate change induced changes in some aspects of the hydrological cycle have not yet been studied in depth. In particular we mention here the contribution from thawing of permafrost.

1.3.2 The contribution from thermal expansion

Jan de Wolde

Many different processes affect sea level change but on the century time scale, thermal expansion of sea water is one of the most important factors that influence sea level, both on the regional and global scale. Thermal expansion relates to the process that the volume of sea water present in the oceans increases if sea water temperatures increase, due to a reduction in sea water density. Changes in the volume of the ocean basins (e.g. due to tectonic or isostatic processes) and in the mass of the ocean water (e.g. due to ice melt) are not taken into account in thermal expansion calculations.

Estimates of thermal expansion are complex, because the density of sea water is a strongly non-linear function of salinity and temperature (Figure 1.15). Variations in the salinity fields are generally small and therefore are often not taken into account in the thermal expansion coefficient. However, the ocean salinity is of great importance in the ocean mixing processes and in the formation of ocean middle and deep water, which determine the ocean thermohaline circulation. Changes in the ocean salinity fields can cause considerable changes in the ocean circulation and thereby in thermal expansion (Manabe and Stouffer, 1994). Since the thermal expansion coefficient increases with temperature, thermal expansion of sea water is not determined by the total amount of heat uptake by the oceans only. Also geographical variations in ocean surface warming and the related heat distribution in the oceans have to be taken into account.

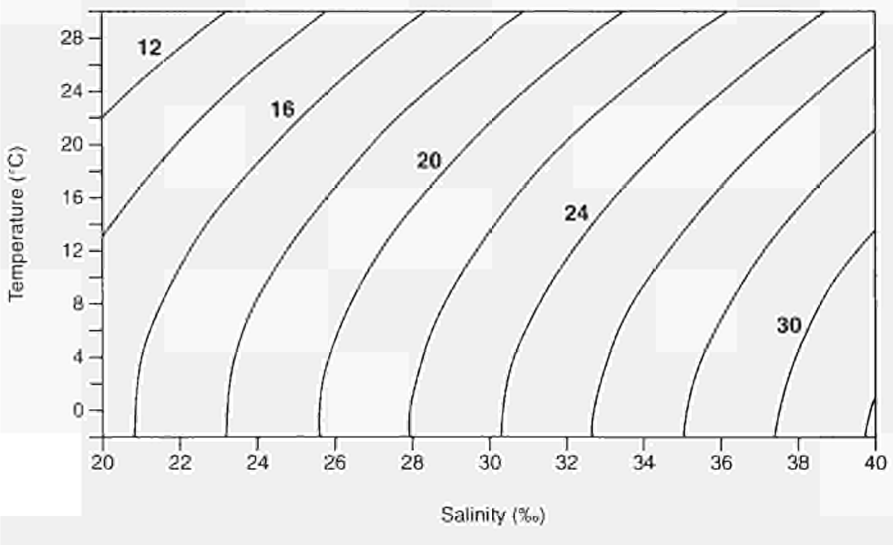


FIGURE 1.15 Density (Kg/m^3-1000) of sea water as a function of salinity and temperature.

If the 3D-thermal structure of the oceans were known as a function of time, changes in the ocean volume due to thermal expansion could be calculated easily. However, observations of the ocean temperature fields are scarce and mostly restricted to the ocean surface only. Some observational studies indicate that the sub-surface ocean waters have warmed in recent decades (Roemmich, 1992; Salinger *et al*, 1996), but conclusions drawn from these studies are limited geographically. Estimates of the contribution of thermal expansion to changes in sea level are therefore mainly based on modelling studies, both for the past and for the future. Since the interaction between atmosphere and ocean plays an important role, only coupled atmosphere/ocean models are applied to estimate thermal expansion.

Until now, most studies on thermal expansion were carried out using 1D upwelling-diffusion models (e.g. Wigley and Raper, 1993; Raper *et al.*, 1996). In these models, the oceans are represented by a globally or hemispherically averaged vertical column of water. The ocean thermohaline circulation is taken into account by prescribing a global mean upwelling velocity (w) and the vertical heat transport is parameterised by diffusion. Because ocean heat mixing processes are still poorly understood, the uncertainties in the diffusion coefficient, k, are large. The vertical temperature profile is determined by the scale-depth (k/w), and another model parameter specifies the change in the temperature of the water that is assumed to downwell outside the model-area. Thermal expansion is calculated from changes in the vertical profile of the area-averaged ocean temperature and results in estimates of the contribution to global mean sea level rise. The largest uncertainty in climate change projections is the climate sensitivity. In 1-D models the climate sensitivity can be a specified parameter, which has advantages for comparison and uncertainty assessment purposes. These models are fast and therefore suitable to perform many scenario runs. Model results can provide insight into the importance of the basic processes included and give information about uncertainties. For the projections of sea level rise presented in the IPCC96 report (Warrick *et al.*, 1996), such a 1D upwelling-diffusion model was used to estimate the contribution of thermal expansion.

A disadvantage of current 1D-models is that several aspects, which can be of interest for thermal expansion, are not or are only poorly represented. Therefore, more recently, 2D (zonally averaged) climate models have been used to investigate these aspects (Harvey, 1994; de Wolde *et al.*, 1997). 2D climate models consist of a zonally averaged single basin ocean model or of a zonally averaged three-basins ocean model coupled to a (zonal mean) energy-balance model of the atmosphere. A schematic illustration is given in Figure 1.16. Ocean heat mixing is parameterised by diffusion, both horizontally and vertically. The ocean circulation is prescribed or calculated from parameterised ocean dynamics (Wright and Stocker, 1991). Since the ocean circulation is treated in a vertical-latitudinal plane, a more physically based ocean heat transport is represented than in 1D-models. Furthermore, 2D-models take into account the variations in thermal expansion coefficient due to latitudinal temperature gradients in the oceans. The changes in temperature of the downwelling polar waters are internally determined in the model. Other aspects, which are represented in 2D-models are the seasonal cycle and variations in sea ice coverage. In the IPCC96 report, some sea level projections are presented that are based on a 2D zonal mean climate model, in order to demonstrate inter-model differences. Due to several aspects, large inter-model differences were found in global mean thermal expansion between 1D and 2D model-results (de Wolde *et al.*, 1997).

Although 2D-models provide some information on the latitudinal variation in thermal expansion, regional variations in sea level change can only be obtained from 3D-models. Changes in the ocean temperature and salinity fields cause changes in the ocean circulation and thereby in thermal expansion, these interactions can only be fully taken into account in 3D-models. Unfortunately, the number of applications of 3D-models to thermal expansion is

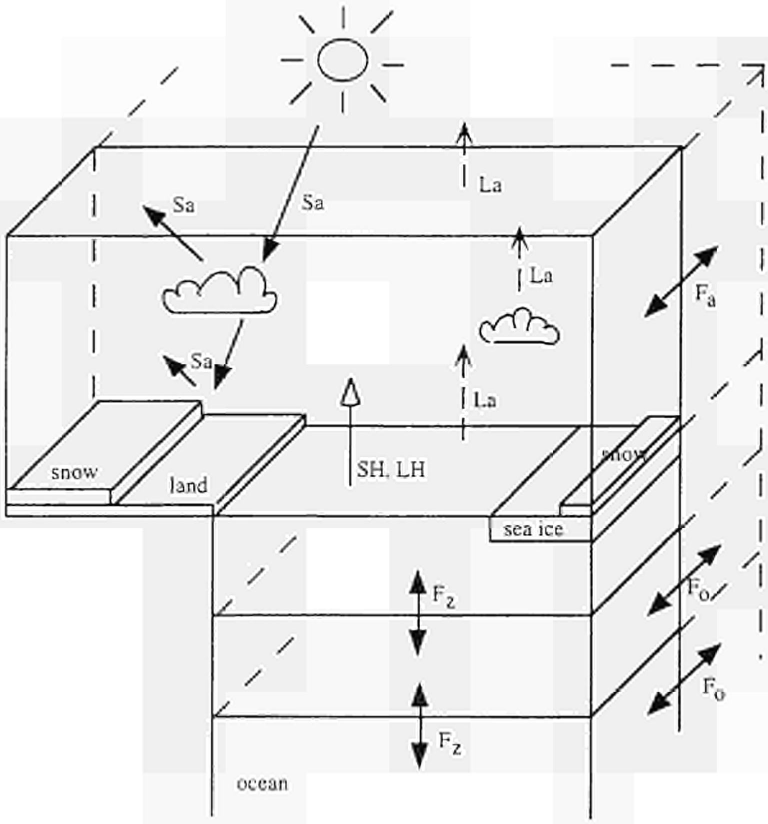


FIGURE 1.16 Schematic illustration (after Harvey, 1988) of a zonal band from the model, containing land, ocean and partial snow and sea ice cover. S_a, L_a, SH and LH represent the shortwave radiation, the longwave radiation and the turbulent fluxes of sensible and latent heat, respectively. F_a and F_O represent the total meridional energy fluxes in the atmosphere and ocean, respectively. F_z represents the vertical heat flux in the ocean

only limited. Church *et al.* (1991) used a subduction ocean model, in which heat transport takes place by advection along isopycnal (constant density) surfaces. But this model is driven by prescribed ocean mixed-layer temperatures, and it ignores the formation of deep water. Atmosphere-Ocean General Circulation Models (AOGCMs) are the most comprehensive tools to study thermal expansion (Mikolajewicz *et al.*, 1990; Cubasch *et al.*, 1992; Gregory, 1993; Manabe *et al.*, 1994). Changes in the model atmosphere and oceans feedback on each other and the ocean circulation is calculated from dynamics. However, AOGCMs are very demanding on computer resources and they often suffer from a climate drift. Most AOGCMs conserve volume rather than mass. Sea level results from AOGCMs indicate that regional variations in sea level change due to thermal expansion are of the same magnitude as the global mean value. But patterns in sea level rise in different AOGCM-results do not always resemble each other, due to the complex interaction between oceans and atmosphere and between ocean circulation and thermal expansion.

Due to a lack of observations of the ocean thermal structure in the past, estimates of the contribution of thermal expansion to observed sea level changes are based on model studies. Raper *et al.* (1996) used an 1D-upwelling diffusion model, which was forced by the estimated global mean radiative forcing changes from increases in greenhouse gas concentrations over the last century. For the period 1880-1990, their estimated range of sea level rise due to thermal expansion is 3.1cm to 5.7cm with a best estimate of 4.3cm. These values are in agreement with the estimated global value of 3.6cm obtained with an AOGCM (Mitchell *et al.*, 1995) that was also forced with historical radiative forcing changes, but taking into account regional variations. De Wolde *et al.* (1995) used all ocean surface temperature data available to construct a regional varying ocean forcing by which their zonal mean ocean model was driven. They obtained an estimate of past thermal expansion varying from 2.2cm to 5.1cm with a best estimate of 3.5 cm. Assuming a uniform temperature increase of the ocean mixed layer temperature in their subduction model, Church *et al.* (1991) estimated 6.9 cm sea level rise due to thermal expansion over the last century. All these estimated values of past thermal expansion are small compared to the observed sea level rise during the last century at many tide gauge stations.

Although estimates of past thermal expansion are in reasonable agreement with each other, projections of (future) sea level rise due to thermal expansion vary more widely. Part of this variation is due to different assumptions about future concentrations of greenhouse gases, which are related to the growth of the world population, world economy, energy demands et cetera. However, even when identical radiative forcing scenarios are assumed, large inter-model differences in projections of thermal expansion are found. For the IS92a-f radiative forcing scenarios (Warrick *et al.*, 1996), the contributions of thermal expansion to global sea level rise in 2100 calculated with the 1D-model (Raper *et al.*, 1996; climate sensitivity value 2.5 °C) range from 21cm to 33cm with a middle estimate of 28cm, whereas these values calculated with a zonal mean climate model (de Wolde *et al.*, 1997) vary from 11cm to 22cm with a middle estimate of 15 cm. Several reasons are identified which are responsible for these inter-model differences (Raper and Cubasch, 1996, de Wolde *et al.*, 1997). Nevertheless, the uncertainties in the projections of thermal expansion due to uncertainties in model parameters are large. In all models, ocean heat mixing processes are parameterised by diffusion, but the magnitude of the diffusion coefficient is not well-known. Figure 1.17 shows the range of uncertainty in thermal expansion calculated by the zonal mean climate model, which is due to the uncertainty in the diffusion coefficient. It can be concluded that the uncertainties in projections of thermal expansion caused by uncertainties in the ocean heat mixing parameters are just as large as the uncertainties introduced by the various radiative forcing scenarios.

Thermal expansion projections from AOGCMs cannot easily be compared to each other, because of different model climate sensitivities and different assumptions concerning emission scenarios and initial conditions. Furthermore, different AOGCMs produce different patterns of sea level change due to thermal expansion. However, all thermal expansion results from AOGCMs indicate that regional variations in sea level change due to thermal expansion are of the same magnitude as the global mean change, because of dynamic effects. In view of this result and in view of the uncertainties in projections of global sea level rise due to thermal expansion, one has to conclude that projections of sea level rise for the European coasts are still largely uncertain. Further efforts should be taken in ocean modelling and ocean monitoring in order to reduce these uncertainties. Observations of the ocean thermal structure are required to enable a better validation of existing ocean models and to improve our knowledge of the ocean heat mixing processes and the heat exchange at the ocean surface. Ongoing ocean modelling is needed to get more insight in the problem of regional variations in sea level changes.

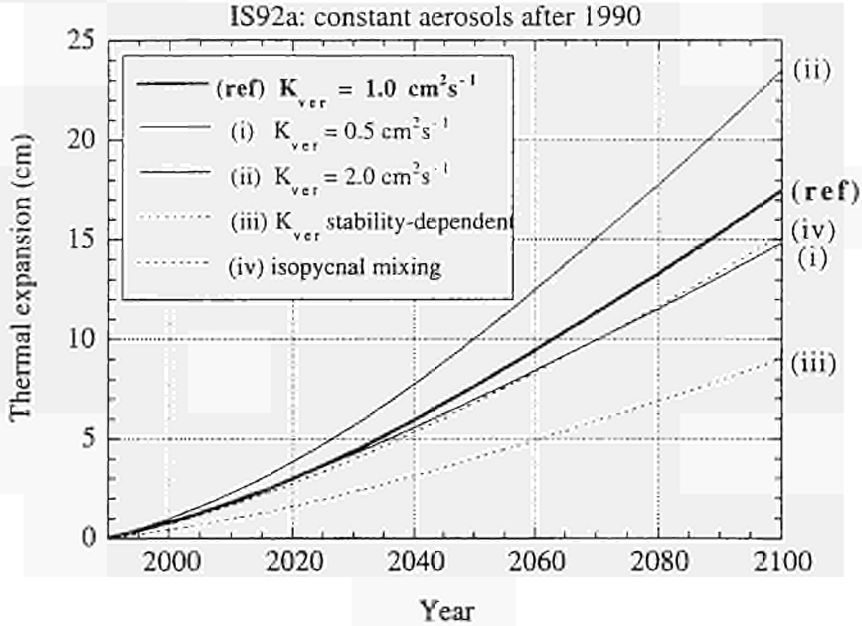


FIGURE 1.17 Sea level contribution of thermal expansion (cm) from 1990 to 2100 for the updated IS92a Scenario with aerosols constant at 1990 level (upper plot): results are shown for the reference case and for four alternative parameterisations of the ocean heat mixing (from de Wolde et al., 1997).

1.3.3 The contribution from glaciers and ice caps

Sarah Raper

1.3.3.1 Introduction. The term *glaciers and ice caps* (GICs) refers to all land-based mountain glaciers, ice fields and ice caps of the world excluding the large ice sheets of Antarctica and Greenland. These range from small ice fields to the large ice caps of, for example, Alaska and Patagonia, where observations are particularly scarce. The locations of the glaciers cover a large range of latitudes, altitudes and degrees of continentality and therefore of climate. Climate change in terms of both temperature and precipitation is also very spatially inhomogeneous. All these factors make estimating the contribution of glaciers and ice caps to sea level change a formidable task.

1.3.3.2 The present volume of ice incorporated in GICs. A prerequisite to estimating past and future contributions of the GICs to ocean volume changes is a knowledge of the present area and volume of the glaciated regions. Both the estimated area and volume of GICs quoted by the IPCC has increased from a 1990 area estimate of 0.55 ±0.05 10^6km^2 and a volume estimate of 0.35 ±0.053 m (equivalent sea level) (Warrick and Oerlemans, 1990) to a 1995 area estimate of 0.68 10^6km^2 (uncertainty estimated at less than 10%) and volume estimate 0.5 ±0.1m (equivalent sea level) (Warrick *et al.*, 1996). These increased estimates are, of course, not due to an actual increase but due to difficulties in estimating these quantities. Estimates of the volume of GICs are still uncertain because the thickness of the ice masses is largely unknown and volume is estimated by extrapolation from the few known area/volume relationships (Meier, 1993, Meier and Bahr, 1996). The present volume of the GICs is clearly important since it sets an upper limit to possible contributions to the ocean volume from this source. The major data source for the estimate of GICs area is the World Glacier Inventory (IAHS ICSI)/UNEP/UNESCO, 1989). Since this inventory covers only 12 regions the data were supplemented using U. S. Geological Survey data for coastal Alaska and adjacent Canada and then extrapolated from the 13 known regions to the 31 regions defined by Meier (1984) according to the character and area of the regions.

1.3.3.3 Changes in individual glaciers over the last 100 years. Significant changes in the volume of individual glaciers over the last few centuries have certainly occurred as shown by documentary evidence (Grove, 1988). Internationally co-ordinated long-term monitoring of glaciers began in 1894, and compilations of data on mass balance, area and length change can be obtained from the World Glacier Monitoring Service (IASH(ICSI)/UNEP/UNESCO, 1993). However, the total number of glaciers on which mass balance was directly and regularly measured was six or less through 1956, fifty to sixty during the period 1967 to 1989, and is now slightly less.

Mass balance, usually measured in m water a^{-1}, is the difference between the accumulation and ablation on the glacier surface. When averaged over the glacier surface cumulative mass balance can give an indication of changes in a glacier's volume with time. In some glaciated regions, however, ice berg calving and internal accumulation (refreezing of melt water within the glacier) are also important (Meier, 1994), though because of difficulties in accounting for these factors they have often been ignored in the global analyses discussed below. Mass balance describes changes on the glacier surface which together with ice flow dynamics lead to changes in the surface profile and area of the glacier. Although numerical ice flow models exist (for example, Greuell, 1992, Oerlemans *et al.* 1998) it is not at present practical to apply such models on a global scale.

Glacier area average mass balance data for individual glaciers has been extended in time using regression analysis on climatological data, usually summer temperature and winter precipitation from a nearby meteorological station. The extended cumulative balance curves show a loss of mass over the past century for glaciers from many parts of the world (see Warrick *et al.*, 1996). A comparison of the regression coefficients for temperature and precipitation together with the range of variation of temperature and precipitation implies that the temperature changes are more important in determining the mass balance (see Chen and Funk, 1990). The mass loss illustrated in Figure 1.18, is therefore generally ascribed to increasing temperatures over this period together with some residual melting occurring due to the climatic warming after the end of the Little Ice Age.

Using a simple geometric model of glacier volume, however, Raper *et al.* (1996) show that although the greater part of the mass loss from Storglaciären during the first half of the century was associated with a climatic warming, the mass loss over the second half of this century is mainly caused by a decrease in accumulation (Figure 1.19). Changes in precipitation may therefore be important in some regions over some periods of time.

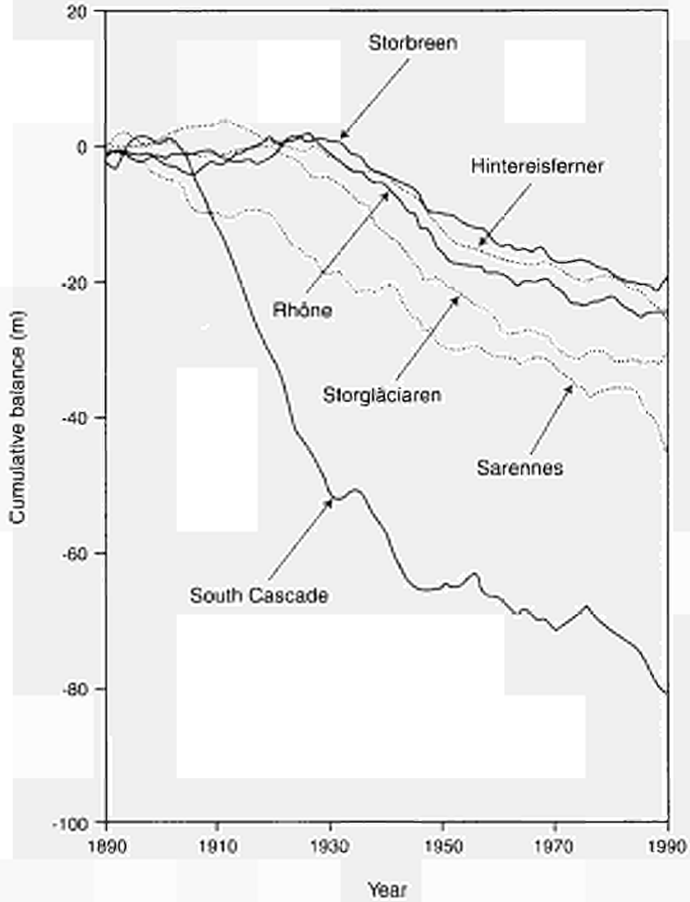


FIGURE 1.18 Cumulative mass balances, in metres of water equivalent, for the glaciers Hintereisferner (Austria), Rhône (Switzerland), Sarennes (France), South Cascade (United States), Storbreen (Norway), and Storglaciären (Sweden). These are among the few glaciers with long observational time series that have been extended using well calibrated hydrometeorological models. All values are relative to 1890. (from Warrick et al., 1996)

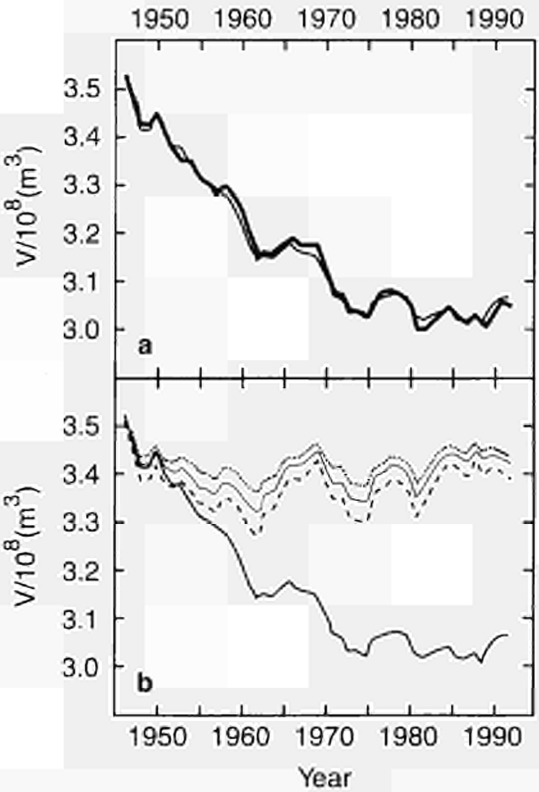


FIGURE 1.19 Observed (thick) and reconstructed volume for Storglaciären, including observed accumulation over 1946-92. (The three reconstructions using different assumptions about the temperature/accumulation correlation are indistinguishable.) (b) Same as (a) but excluding observed accumulation over 1946-92. The reconstructions commenced in AD 500. (from Raper et al., 1996)

1.3.3.4 Global scale observation based changes over the last 100 years. Meier's (1984) estimate of the contribution of GICs to global sea level for the period 1900 to 1961 is 0.46 ±0.26 mm a^{-1}, updated to 0.40 ±0.2 mm a^{-1} in Meier (1993). It is still perhaps the best observation based estimate for the past century. These estimates are based on mass balance and volume change data for 25 glaciers representing 13 glaciated regions of the world. Data gaps for the period were filled in using hydrometeorological models. To extrapolate these results to cover the 25 glaciated regions of the world, Meier assumed that the magnitude of the long-term mass balance could be related to the magnitude of the annual mass balance amplitude, that is the mean of the positive sum of the winter and summer balances. The annual amplitude is a measure of the mass turnover and depends mainly on the climatological regime. For example, it is highest in temperate maritime regions and decreases towards the poles and with increasing continentality. However, the correlation between the long-term mass balances and the annual amplitude for the 13 regions with data is only -0.55. This rather low correlation gives rise to uncertainties which fall within the quoted uncertainty range. Meier's results include an adjustment for the decrease in glacier area between 1900 and 1961. The product of the annual amplitude multiplied by the area of the glaciated region is a measure of the contribution of that region. From Meier's calculations one third of the mass contribution comes from Gulf of Alaska, and large contributions also come from Central Asia and Patagonia all regions underrepresented by the data. The updated estimate is smaller than the original mainly due to allowance for internal accumulation.

In an independent analysis Schwitter and Raymond (1993) estimate a mean rate of contribution to the global ocean volume from GSICs of about 0.5mm a^{-1} since the last Neoglacial. This estimate has large uncertainties but agrees well with Meier's estimates discussed above. Schwitter and Raymond's (1993) estimate is based on glacier profile changes from the Neoglacial (estimated from geomorphological markers) to the present (sequential map information and surveyed profiles). A profile-shape factor, f, defined as the ratio of the average thickness change along the length of the glacier to the thickness change at the present terminus was evaluated. A value of f of 0.5 would assume a linear decrease in thickness change from the terminus to the head of the glacier; the 15 mountain glaciers studied give an average value of f of 0.28 The volume change above the present terminus is estimated as $f\Delta h(l_0)a_0$, where a_0 is the area above the terminus and $\Delta h(l_0)$ is the thickness change at the terminus. The uncertainties are large, for example, the volume lost below the present terminus is not included but is estimated to imply a correction of about +20% for the largest glaciers and +50% for the small glaciers. However, Schwitter and Raymond note that the uncertainties in f and $\Delta h(l_0)$ are also large and the extrapolation of the results based on small to medium sized glaciers to the large glaciers of coastal Alaska and sub-polar ice caps is questionable.

1.3.3.5 Modelling the sensitivity of glaciers to climate change. For modelling the sensitivity of individual glaciers to climate change more detailed mass balance models, than the glacier area mean mass balance models based on simple regression against temperature and precipitation variables as discussed above, have been developed. There are two main types of more detailed mass balance model, both of which seek to model the mass balance profile over the altitudinal range of the glacier. These are energy balance models (Oerlemans and Fortuin, 1992) and degree-day models (Laumann and Reeh, 1993, Jóhannesson *et al.*, 1995). These two types of model give an estimate of the static sensitivity to a small change in a climatic variable such as temperature. As well as estimating the static sensitivity such models can be used to force ice-dynamic models and hence estimate the response of a glacier to climate change. However, as mentioned earlier such coupled models have not yet been used to estimate the response of the glaciers to climate change on a global scale.

Oerlemans and Fortuin (1992) use an energy balance model to fit the mass balance profiles of 12 glaciers which are representative of a wide range of climatological settings. The mass balance model is based on the equation:

$$B = \int_{year} [(1-f)\min(0;-\psi L) + P^*]dt \qquad (1)$$

where L is the latent heat of melt, ψ is the energy balance at the ice or snow surface, and f is the fraction of melt water that does not run off but refreezes. P^* is the rate of solid precipitation. Thus the first term represents the melt and runoff due to a positive energy balance and the second term represents the accumulation. The energy balance depends on the solar radiation, atmospheric radiation, turbulent fluxes of heat and moisture and the energy used for heating up the upper snow or ice layers (Oerlemans, 1993). The albedo was calculated in the model so that the albedo feedback is included. The model input was climatological information and the altitudinal gradients in precipitation were tuned to give a good match between the observed and modelled mass balances. Mass balances averaged over the glaciers were obtained by area averaging the mass balances profiles. Changes in these mean specific balances for the 12 glaciers for a 1 °C warming were in the range -0.12 to -1.15 m a^{-1}.

Oerlemans and Fortuin (1992) found that the sensitivity of glacier mass balance to a 1°C warming increased logarithmically with the glacier precipitation rate (Figure 1.20). This finding supports Meier's assumption that the sensitivity is proportional to the mass turnover. Oerlemans and Fortuin showed how this precipitation versus sensitivity relationship could be used to extend the analysis to the other glaciated regions of the world. Using 100 glaciated regions with an assigned value of annual precipitation, the area weighted mean change in the specific mass balance for a uniform 1 °C warming is -0.395 m a^{-1}. However, it is not possible to meaningfully compare this value with Meier's estimate of melt over the past century because the warming over the globe has clearly not been uniform. Using regional temperature change data Zuo and Oerlemans (1997) estimate the loss of ice from GICs contributes 2.7±0.9 cm to sea level over the period 1865-1990. The uncertainty reflects uncertainty in the initial glacier state with respect to climate. This estimate is based on the static sensitivity to a 1 °C warming and does not account for changing glacier areas.

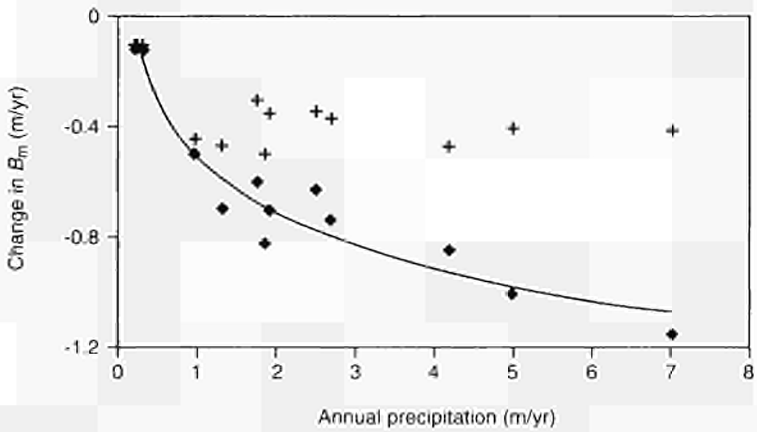


FIGURE 1.20 Effect of a 1°C warming on the mass balance of the 12 selected glaciers, plotted in dependence of the annual mean precipitation. Bm is the mean-specific balance. Crosses refer to an experiment where summer temperature only was raised by 1°C (from Oerlemans and Fortuin, 1992).

Degree-day mass balance models have the advantage that they are formulated in terms of temperature and precipitation, variables which are more readily available than those needed for energy balance models and which are typically given in climate change scenarios. Laumann and Reeh (1993) applied a degree day model, calibrated on climatological 30 year normals of temperature and precipitation, to three glaciers in a west-east transect in southern Norway. Again this study indicated that the low-lying maritime, high-precipitation glacier in the west is more sensitive to climate changes than the dryer more continental higher glaciers in the east. Changes in the specific mass balance for the three glaciers studied ranged from -0.54 to -1.04 m a^{-1} for a 1 °C warming. Laumann and Reeh (1993) also concluded that an increase in precipitation of 20-40% per °C warming would be needed to offset the effect of the warming on melt. The greater sensitivity of a maritime glacier compared to a more continental Arctic glacier is illustrated in Figure 1.21. Even allowing for the different altitudinal distribution of the area of a valley glacier such as Ålfotbreen compared with the Devon Island ice cap the higher static sensitivity of Ålfotbreen is clear.

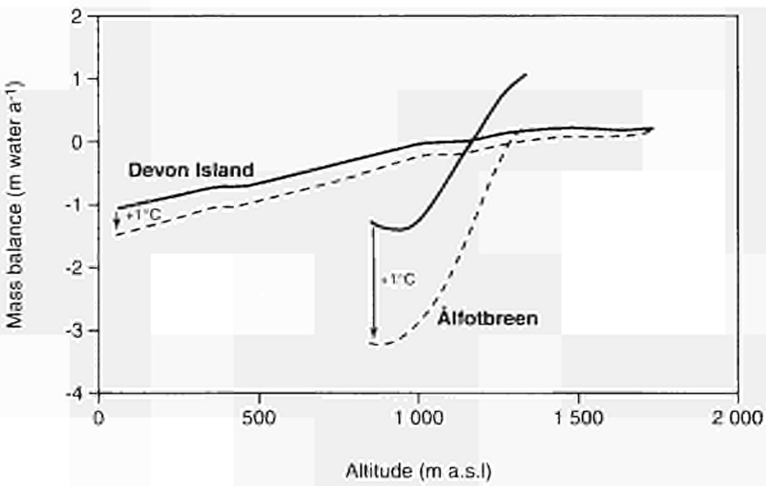


FIGURE 1.21 The low static sensitivity of an Arctic glacier (Devon Island ice cap, N.W.T., Canada) compared with the high sensitivity of a very maritime glacier (Ålfotbreen, western Norway)(from R.J. Braithwaite unpublished).

Degree-day mass balance models determine the snow accumulation, melting of snow and ice, and refreezing as a function of altitude based on observed temperature and precipitation at a meteorological station (Jóhannesson *et al.*, 1995). Temperature on the glacier is estimated assuming a constant lapse rate. Precipitation may be estimated assuming a linear precipitation variation with altitude as in the above reference, or alternatively, precipitation may be used as a tuning parameter so that the observed and modelled altitudinal distribution of mass balance coincide. The altitudinal variation of snow accumulation is determined from the precipitation and temperature with a typical threshold for snow versus rain of 1°C. When the precipitation and/or temperature is specified as annual or monthly means a statistical approach is used to determine the amount of precipitation that falls as snow and the number of positive degree-days following Braithwaite (1984). Melting of snow and ice is calculated from the number of positive degree-days depending on the estimated degree-day factors for snow and ice for that particular glacier. Allowance is made for the refreezing of a specified fraction of the snow melt. The mass balance at each altitude is the sum of the ablation and accumulation over a time interval.

Jóhannesson *et al.* (1995) applied a degree-day mass balance model to three glaciers in Iceland, Norway and Greenland using daily and monthly precipitation and temperature data instead of climatological annual mean data. Model calibration was carried out using annual accumulation, ablation and mass balance measurements as opposed to long-term averages. The model parameters were found to be stable with time implying the model is suitable for studying climate change scenarios.

1.3.3.6 Modelling future changes in the global glacier volume. Two models were used by the IPCC in the second assessment to make future projections of the contribution of GICs to the ocean volume. For the calculations starting from the one-dimensional upwelling-diffusion model, the GIC calculation used a simple semi-empirical model based on Raper *et al.* (1996), which relates glacier volume changes to global mean temperature change (Wigley and Raper, 1995). There are three important parameters in this model: (i) the initial (1880) global ice volume, which was assumed to be 30 cm sea level equivalent (low compared to the recent estimate by Meier (1993) of the present volume at 50 cm sea level equivalent); (ii) the minimum temperature increase which, if it were maintained, would cause a given glacier to eventually disappear; and (iii) the glacier response time. The model equation for an individual glacier is

$$dV/dt = (V - V_\infty) / \tau \qquad (2)$$

where t is time, V is the annual mean glacier volume, V_∞ is the equilibrium value of V assumed to be linearly dependant on temperature, and τ is the glacier response time. Because there is a distribution of critical temperature increase and glacier response times in nature, the calculations assume a distribution of minimum temperature increases required for disappearance of a glacier, and of glacier response times. The model was calibrated by choosing parameters which produce model-based estimates of past GIC volume change which are consistent with observationally-based estimates. Uncertainties arise because the problem is underdetermined. Using observed global and annual mean temperature changes over 1900 to 1961 the parameters were adjusted to give a sea level contribution of 1.6 cm with an uncertainty range of ±50%. The central value chosen is somewhat lower than the estimate of Meier (1993) vis 2.4 cm. The results over the past century are sensitive to the base year temperature which determines the initial balance, but for the future because of compensating influences the effect becomes smaller. Though there is clearly scope for improvement, this model has the advantage that it allows for the diminishing volume of the GICs.

For the calculations starting from the two-dimensional upwelling-diffusion model (de Wolde, *et al.*, 1995), the sensitivity of glacier mass balance to changes in temperature is assumed to be latitudinally dependant based on the mean annual precipitation of the 100 glacier regions following Oerlemans and Fortuin (1992). Latitudinally and seasonally varying change in the surface air temperature from the model were used as input to the ice melt model. Changes in glacier area are not taken into account. It was hoped that on the time-scale considered the warming associated with the lowering of the ice surface and the decrease of the area due to the retreat of the glacier terminus will tend to have a counterbalancing effect on the sensitivity values. Model calculations start in 1990, although at present most glaciers are not in equilibrium. To account for the observed present-day thinning of glaciers, a constant long-term trend of 0.5mm a^{-1} sea level rise was added to the projections.

Despite the very different approaches of the two models used by IPCC for the projections of GIC melt, they both give similar results for the next century. For the future IPCC scenario IS95a (including aerosol changes), the mid estimate for the contribution to sea level rise over 1995-2100 from the 1-D approach is 16 cm compared with 12 cm from the 2-D approach.

Because of the shortcomings in both approaches, however, this agreement is probably largely coincidental. More recently, a calculation based on regional and seasonal temperature change from an A/OGCM gives rise to a larger glacier melt by 20% compared to that using annual and global mean temperature change (Gregory and Oerlemans, 1998).

1.3.4 The contribution from the Greenland ice sheet: reduction of uncertainties

Niels Reeh

1.3.4.1 Introduction. There is a general consensus that the glaciers and ice caps of the Earth have contributed to sea level rise over the last 100 years and will continue to do so at an increasing rate in a warmer climate (Warrick *et al.*, 1996). As regards the large ice sheets in Greenland and Antarctica, the situation is different, because it is not possible to say with any certainty whether the ice sheets are currently in balance or have increased or decreased in volume over the last 100 years. The uncertainty about the present state of balance of the Greenland and Antarctic ice sheets is of course inherent in predictions of future contributions to sea level from the two big ice sheets. The general belief is that increased melt rates at the margins of the Greenland ice sheet in a warmer climate should dominate over any increase in accumulation rates in the interior, potentially causing sea level to rise. With respect to the Antarctic ice sheet, a warmer climate should increase accumulation rates and thus potentially cause sea level to fall. Thus, the future contribution to sea level change from the ice sheets in Greenland and Antarctica is the sum of the current (background) contribution, for which not even the sign is known and the future climate related potential rise/fall (whether caused by anthropogenic or natural climate change).

The large uncertainty about the present state of balance of the Greenland and Antarctic ice sheets is due to the fact that neither mass balance components (snow accumulation, ice melt and iceberg calving) nor surface elevation change have yet been observed over a sufficiently long period of time to provide reliable estimates of the long-term trend in ice volume change.

The main purpose here is to discuss the possibility of reducing the uncertainty of the estimate of the present as well as the future contribution to sea level from the Greenland ice sheet. However, much of the discussion is equally relevant for the Antarctic ice sheet and part of the discussion is even relevant also for the Earth's glaciers and ice caps.

The emphasis is on the long-term (century scale) background contribution from the ice sheet to sea level change, which we must know in order to reliably predict the future contribution, but which, at present, is largely unknown. As regards the contribution to sea level change caused by future climate change, the main uncertainty is not so much on the glaciological side, but is due rather to the difficulty of predicting future climate change on regional scales. With a given climate scenario, the future climate change contribution to sea level from the ice sheets can be estimated much more accurately than the long term background contribution, as a result of recent studies of ice-sheet mass-balance sensitivity to climate change carried out in several CEC funded projects (CEC, 1993;1995). These studies provided new knowledge about both spatial and temporal variations of ice sheet surface albedo (e.g. Zuo and Oerlemans, 1996), and degree day factors (Braithwaite, 1995) that are important input parameters to mass balance/climate sensitivity models. Moreover, the mass balance studies were extended to cover ice margins in North-East and North Greenland where essentially no measurements had previously been carried out (Bøggild *et al.*, 1994; Konzelmann and Braithwaite, 1995). Also the question of ice margin (outlet glacier) instability was addressed by the detection and study of a major surge of Storstrømmen, a large outlet glacier from the North-East Greenland ice-sheet. The surge and post-surge phases are documented by Reeh *et al.* (1994).

The studies of ice-margin mass-balance and stability in North-East Greenland continue as part of the on-going CEC funded project 'Climate Change and Sea Level'. An important purpose of these studies is to measure, for the first time ever, the bottom melting from an extended floating glacier tongue in Greenland (Thomsen *et al.*, 1997). Preliminary results indicate that bottom melting from North and North-East Greenland floating glacier tongues may presently constitute as much as 5% of the total mass loss from the Greenland ice sheet, a hitherto overlooked term.

1.3.4.2 Estimated contributions to global mean sea level change over the last 100 years. Some physical characteristics of the current land ice masses, and the contributions to sea level rise over the last 100 years as estimated by Warrick *et al.* (1996) are given in Table 1.3 and Table 1.4, respectively.

TABLE 1.3 Some physical characteristics of ice on Earth. (Modified after Warrick et al., 1996)

	Antarctic Ice Sheet (grounded ice only)	**Greenland Ice Sheet**	**Glaciers and Ice Caps**
Area (10^6 km^2)	12.1	1.71	0.68
Volume (10^6 km^3)	29	2.95	0.18 ± 0.04
Volume (sea level equiv. m)	73 (68)	7.4	0.5 ± 0.1
Accumulation (km^3 of ice equiv. a^{-1})	1810	603	731 ± 100
Accumulation (sea level equiv. m a^{-1})	0.0045	0.0015	0.0018
Change in sea level equivalent for a 10% imbalance (mm a^{-1})	0.45	0.15	0.18

TABLE 1.4 Estimated contributions to global mean sea level rise over the last 100 years in cm. (After Warrick et al., 1996)

Component	**Low**	**Middle**	**High**
1. Thermal expansion	2	4	7
2. Glaciers and ice caps	2	3.5	5
3. Greenland ice sheet	-4	0	4
4. Antarctic ice sheet	-14	0	14
5. Surface/ground water	-5	0.5	7
6. TOTAL	-19	8	37
7. Estimate from tide gauge data	10	18	25

The "zero" entries in Table 1.4 for the Greenland and Antarctic ice sheets should be interpreted as a reflection of the current poor state of knowledge, rather than as an estimate of the current state of balance of the ice sheets. Also the wide range of uncertainty indicated in the table for the Greenland and Antarctic ice sheets is rather arbitrarily set as corresponding to an imbalance of 25% of the annual mass turnover.

We have shown in section 1.2 that the accurate estimation of sea level change for the last century from the tide gauge records is problematical. The same may be said at this time for the estimation of the present mass balance of the ice sheets. However, the more accurate estimation of either could be used to put limits on the other, assuming that the rest of the contributors were well known. Although we are not currently in a position to draw any conclusions from such an exercise we make the calculation here because it illustrates the extent of our uncertainty about the mass balances of Greenland and Antarctica. So the numbers in Table 2 could be used to put limits on the combined contribution from the Greenland and Antarctic ice sheets, if we accept, for the moment, the tide gauge based estimate of sea level rise over the last hundred years of between 10 and 25 cm, and that the contributions to this rise from thermal expansion, glaciers and ice caps, and surface/ground water sum up to be in the range -1 to +19 cm. Using these numbers, the combined contribution to sea level change over the last 100 years from the Greenland and Antarctic ice sheets would be between -9 and +26 cm, as compared with the range (-18 to +18 cm) obtained directly from the low and high estimates for the Greenland and Antarctic ice sheets given in Table 1. This would show that the Greenland and Antarctic ice sheets together are more likely to have caused a sea level rise than a fall over the last hundred years, although a fall of up to 9 cm could not be excluded.

The corresponding range of ice volume change is from + 360 to -1040 km^3 of ice equiv. a^{-1}. If the change were ascribed solely to the Greenland ice sheet, these range limits correspond to a total elevation change over the last hundred years of between +21m and -61 m averaged over the entire ice sheet area. If, on the other hand, the ice volume change were ascribed solely to the Antarctic ice sheet, the total area-averaged elevation change over the last 100 years, would have been in the range from +3 m to -8.6 m. It is hardly realistic to think that either of the two ice sheets has been so much out of balance, although there is not any direct evidence against it.

1.3.4.3 The mass balance equation. The basic equation, expressing the law of mass conservation for an ice sheet in total or for a region of an ice sheet may be written (Figure 1.22):

$$\partial V/\partial t = Q_a - Q_m - Q_c + Q_B. \tag{3}$$

In Equation (3), V is the ice volume, t is time, Q_a is the annual surface accumulation, Q_m is the annual loss by surface melting, Q_c is the annual loss by calving of icebergs, and Q_B is the annual balance at the bottom (melting or freeze-on of ice). All volumes are expressed in terms of ice equivalents.

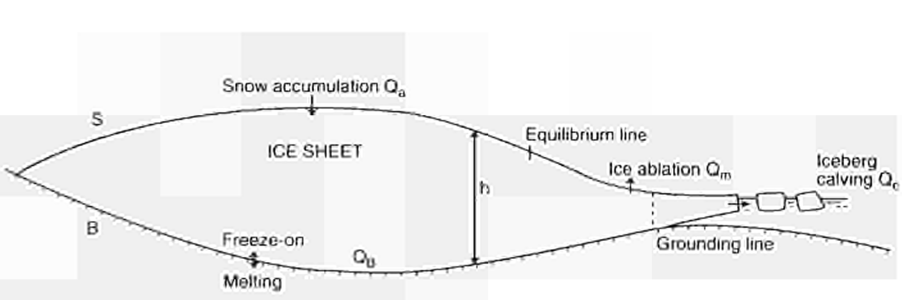


FIGURE 1.22 Ice sheet mass-balance terms.

Equation (3) suggests that the total mass balance of an ice sheet can be determined by two entirely different methods:

(1) Direct measurement of the change in volume by monitoring surface elevation change.

(2) The budget method, by which each term on the right hand side of the mass balance equation is determined separately.

A common problem with both methods is that all terms in equation (3) display large temporal fluctuations on a year-to-year and even on a decadal time scale. So even if the terms could be measured accurately over several years, the measured or calculated volume change might just reflect a statistical fluctuation and not the long-term trend needed for predicting the background contribution from the ice sheet to sea level change. In order to determine this long-term trend, mean values over a period on the order of a century are needed.

The observed elevation change along the EGIG line across the central Greenland ice sheet, and our present knowledge about the accumulation distribution in Greenland, will be used to illustrate these problems.

1.3.4.4 Surface elevation change along the EGIG line. Surface elevation along the EGIG line across the central Greenland ice sheet (Figure 1.23) was measured by using trigonometric levelling in 1959, 1968 and 1992. In the 9 year period from 1959 to 1968, the centre part of the profile experienced a mean increase in surface elevation of +9 cm a^{-1} (Seckel, 1977). However, in the subsequent 24 year period from 1968 to 1992, the same part of the profile experienced a mean surface lowering of -15 cm a^{-1} (Kock, 1993; Möller, 1996), see Figure 1.24.

This shows that, although on-going remote sensing studies based on space-borne radar altimetry (Davis *et al.*, 1998) or airborne laser altimetry (Krabill *et al*, 1995) have the potential to provide a dense spatial coverage of surface elevation data with decimetre or even centimetre accuracy, short term (decadal-scale) fluctuations in surface elevation, probably caused by decadal-scale fluctuations of snow accumulation, are so large that, even if such observation programmes were run over several decades, they are not likely to give us the long-term trend in ice volume change required for determining the current background contribution from the large ice sheets to sea level change.

1.3.4.5 Accumulation distribution in Greenland. The estimate of the total annual surface accumulation on the Greenland ice sheet is based on snow-pit and firn-core studies supplemented by precipitation data from climate stations in the coastal region. The most recent compilation of such data by Ohmura and Reeh (1991) used information from 251 pits and cores spread over the ice sheet area, retrieved in the 77 year period from 1913 to 1990 (Figure 1.26). The accumulation rate at each of the 251 locations is determined as the mean of a varying number of annual values (typically 5-10). The precipitation values for the coastal sites are determined similarly, except that, in most cases, longer record lengths allow longer averaging periods (up to 30 years) to be used.

Inspection of the distribution of observation points in Figure 1.25 reveals that the map is made from a spatially inhomogeneous data set. However, the temporal inhomogeneity of the data set is probably an even larger problem. No attempt was made to reduce the accumulation values to a common standard period. According to Braithwaite (1994) the mean coefficient of variation of annual precipitation values in Greenland is 0.26 ± 0.06, indicating that a 5-10 year mean accumulation-rate value can easily deviate 10 % or more from the long term (century scale) mean value. In spite of these deficiencies, there is a tendency to consider

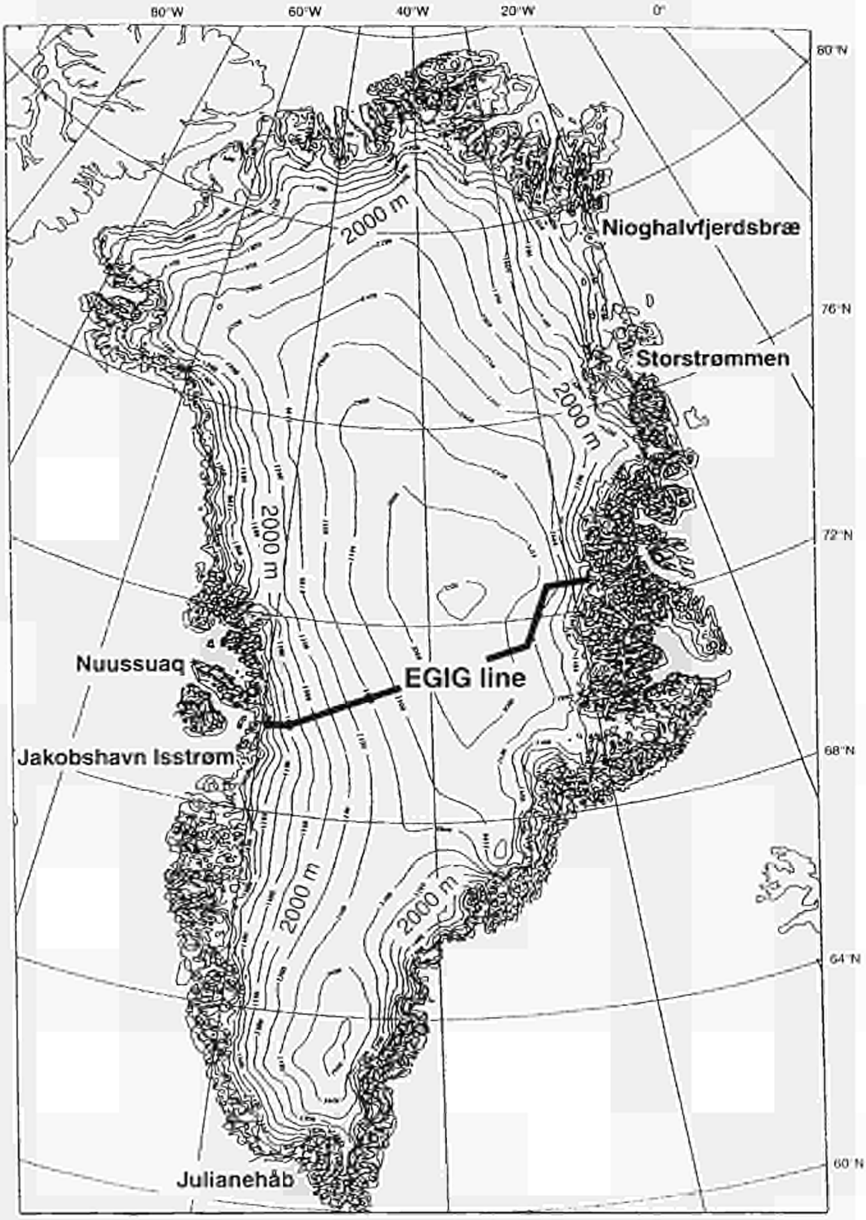


FIGURE 1.23 Map of Greenland showing locations mentioned in the text.

the map shown in Figure 1.25 as the "truth", against which to validate the performance of other methods for estimating accumulation rates, such as using passive microwaves (e.g. Zwally and Giovinetto, 1995) or GCM modelling (Genthon, 1994a; Ohmura and Wild, 1995). The use of the map in Figure 1.23 as a "check-list" is justified only by the fact that no better map is available, and certainly not by its accuracy.

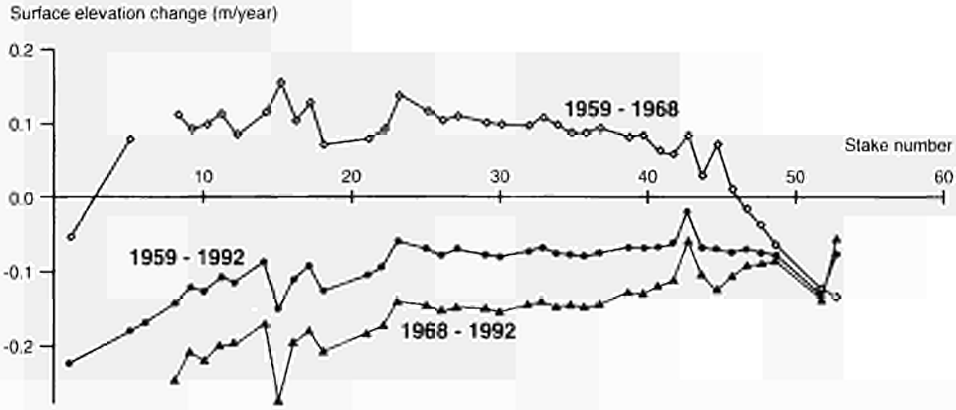


FIGURE 1.24 Surface elevation change along the EGIG line as determined by repeated trigonometric levelling in 1959, 1968 and 1992. (After Kock, 1993).

1.3.4.6 Suggestions for new mass balance studies. It is, however, possible to point at studies that, within a period of a few years, are likely to significantly reduce the present uncertainty on the long-term mass-balance trend of the Greenland ice sheet.

Mapping the trim line zone.
The maximum extent in historical times of the Greenland ice sheet margin, corresponding approximately to the extent 100 years ago (Weidick, 1995) is marked by a distinct trim line zone. Weidick (1968) measured the height of the trim line along glacier lobes at the West Greenland ice sheet margin from Julianehåb district in the south to Nussuaq peninsula in the north (for locations, see Figure 1.24). Based on these observations, Weidick (1968) calculated the mean annual thinning of the ice margin in West Greenland to be approximately 0.7 m a^{-1}. The total annual volume loss from the ice sheet margin, assuming that the whole marginal area of the ice sheet had been subject to the same loss as West Greenland, was calculated to 200 km^3 of ice equiv. per year, a loss that, if maintained over a 100 year period, would result in a sea level rise of 5 cm, which is well inside the possible range discussed previously.

Weidick's study should be repeated and extended to the ablation zone around the whole Greenland ice sheet by using modern mapping methods such as, for example, interferometric SAR supported by tie points provided by *In Situ* GPS measurements and/or airborne laser altimetry measurements. The launch of planned new satellite sensors like ASTER (Advanced Spaceborne Thermal Emission and reflection Radiometer) and GLAS (Geoscience Laser Altimeter System) will, if realised, provide enhanced tools for conducting this research.

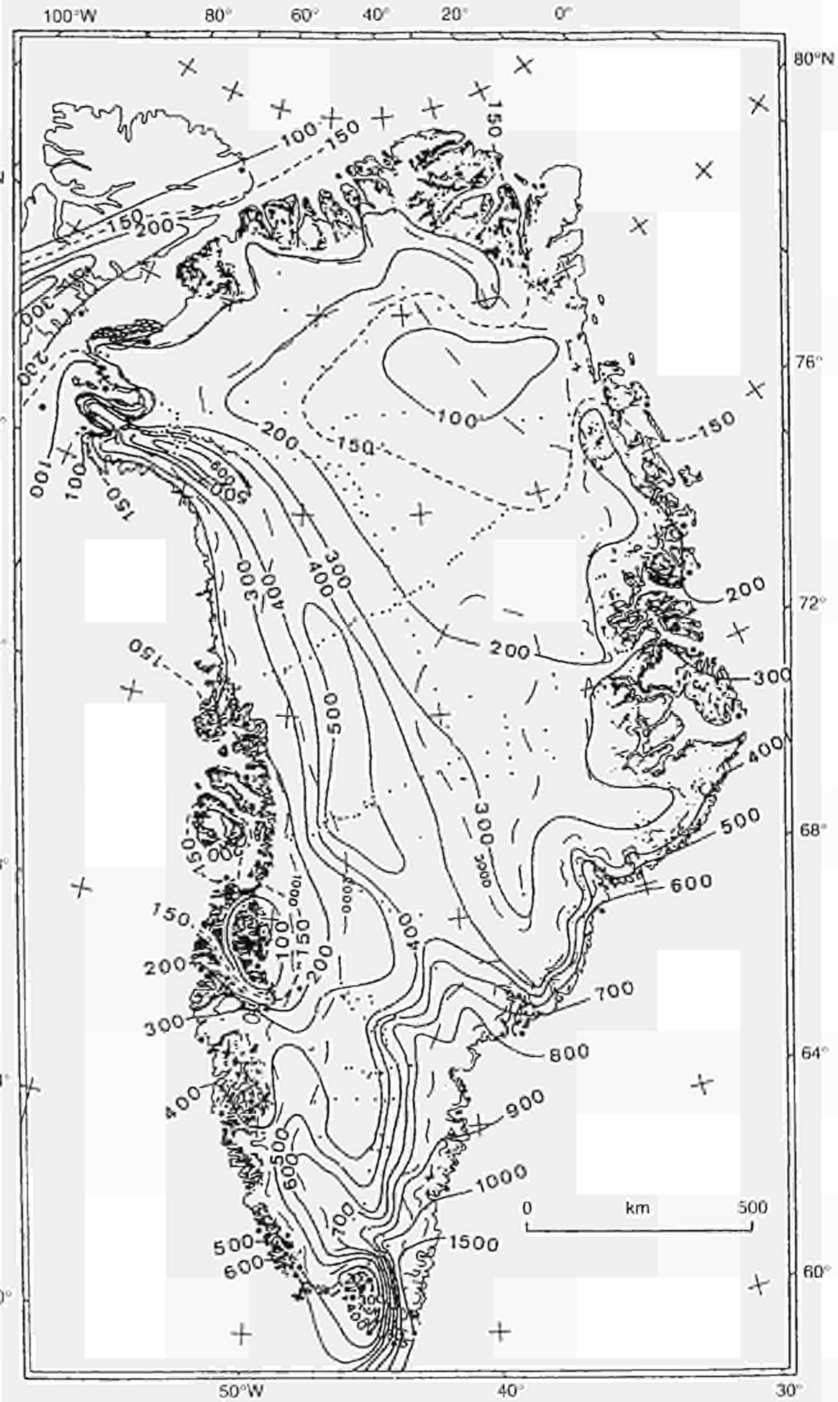


FIGURE 1.25 Snow accumulation in Greenland in mm H_2O/year. (From Ohmura and Reeb, 1991).

Accumulation rate survey by using high resolution radio-echo sounding

However, to estimate the long-term trend of ice sheet volume, it is not sufficient to deal with past changes in the marginal zone, since a volume loss there could very well have been compensated - partly or fully - by a gain in the ice sheet interior region. Unfortunately, there are no reliable observations of surface elevation change for the interior covering the last 100 years. However, the 2000 metre velocity traverse experiment (Thomas and Csatho, 1996), aiming at measuring the ice flux along the 2000 metre elevation contour line around the whole Greenland ice sheet (Figure 1.24), is likely to be only a little influenced by short-term variations, because in the ice sheet interior, quantities that determine ice flow (e.g. surface slope, ice thickness, internal temperature, basal conditions) are likely to show little variation on a century time scale. Streams of fast flowing ice embedded in the ice sheet, for example the ice stream feeding the Jakobshavn Isstrøm in West Greenland (Echelmeyer and Harrison, 1990) or the ice stream feeding several large Northeast Greenland outlet glaciers (Fahnestock *et al.*, 1993), may, however, cause significant short term changes of ice dynamics also in the interior part of the ice sheet.

Apart from possible problems caused by potentially unsteady ice stream flow, and the uncertainty related to estimating the depth averaged velocity from the velocity at the surface, the velocity data along the 2000 metre elevation contour line, when combined with airborne radar ice thickness observations, should give a reasonable estimate of the long-term average mass flux out of the central region of the Greenland ice sheet.

For assessing the mass balance of the region, a good spatial coverage of long-term mean accumulation rates is required. Today, such long-term mean accumulation rates are only available at some 15 drilling locations on the Greenland ice sheet. However, internal horizons of equal age in the upper layers of the ice sheet (so called isochrones) detectable by means of high resolution radio-echo sounding can be dated by correlation with acid layers found in ice cores (Hempel *et al.*, 1993). Accounting for the low-density snow and firn near the surface and correcting for thinning of the layers due to flow deformation, the depth to the reflecting isochrones can be translated into average accumulation rates over the past decades, centuries, or even millennia. Comparing the mass input calculated by means of such radio-echo-sounder derived long-term mean accumulation rate distributions with the mass flux out of the 2000 metre elevation contour line should give an estimate of the long-term "background" mass change of the central Greenland ice sheet.

Together with the mass change over the past c.100 years determined for the ablation region by the "Weidick-method", this should give us a more reliable estimate of the long-term background contribution from the total Greenland ice sheet to sea level than can be obtained by other observation methods.

Modelling

Considering the very long time scales involved in ice sheet evolution, it seems unlikely that the Greenland ice sheet has adjusted completely to its past mass balance history, as demonstrated by a model study of Huybrechts (1994a), see Figure 1.26. Huybrechts (1994a) simulated the evolution of the ice sheet during the last glacial-interglacial cycle, and subsequently derived local thickness changes for the present time. The model results displayed in Figure 1.26 are not necessarily comparable with regional (Zwally *et al.*,1989; Weidick, 1991; Davis *et al.*, 1998) and local (Reeh and Gundestrup, 1985) surface elevation changes derived from observational data, because of different time scales (decadal means for the observational data versus 200 year mean values for the model). Nevertheless, in particular with respect to the sign of the evolution, there is a reasonable agreement between model results and observations.

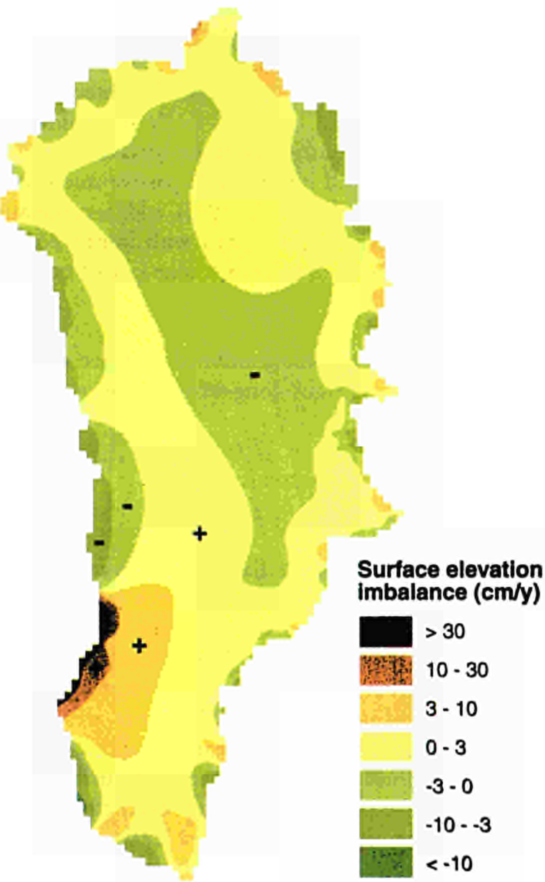


FIGURE 1.26 Present surface elevation imbalance of the Greenland ice sheet: rate of surface elevation change (cm/year) averaged over the last 200 years as obtained after modelling the last glacial-interglacial cycle. Positive values indicate thickening, negative values indicate thinning as marked by the respective symbols (from Huybrechts, 1994a).

Figure 1.26 indicates that large spatial differences of surface elevation change can be expected. There are many reasons for these differences. To mention a few, they could be caused by regional differences in past climate history, or by regional differences in accumulation rate and flow dynamics leading to different warming rates of the basal ice layer.

Figure 1.26 suggests that the present rate of change of ice sheet volume or mass has a component that is determined by pre-anthropogenic forcings operating on century or millennial time scales. We can probably measure this component by means of the observational procedures mentioned above. However, in order to understand the mechanisms behind the current rate of change of ice sheet volume, which is important for predicting future changes, modelling of the ice sheet evolution over the last glacial/interglacial cycle (the last 100,000 years) is required. This emphasises the need for investigating past climate forcings over such long periods (to be obtained e.g. from deep ice core records). But also studies of present ice sheet dynamics and mass balance are needed in order to develop more realistic models for ice sheet evolution.

Space- or airborne remote sensing techniques (such as SAR interferometry) for simultaneous determination of glacier topography and surface velocity (Joughin *et al.*, 1996; Mohr *et al.*, 1998), radar- or laser altimetry for measuring accurate surface elevations (Davis *et al.*, 1998; Krabill *et al.*, 1995), and visible, thermal infrared and passive microwave data for estimating surface temperature and albedo (Steffen *et al.*, 1993) and accumulation rate (Zwally and Giovinetto, 1995) which - at an increasing rate and with improved area coverage - provide data relevant for glacier and ice sheet mass balance studies, are of utmost value for developing improved ice sheet models.

In conclusion, remote sensing observations, combined with ice sheet modelling may be an alternative way of obtaining the current long-term background contribution from the Greenland ice sheet to sea level change.

1.3.5 The contribution from the Antarctic ice sheet

Massimo Frezzotti

1.3.5.1 Introduction. More than 87% of the Earth's fresh water presently exists in a frozen state and more than 90% of that ice occurs in Antarctica (Meier, 1983). The Antarctic ice sheet covers an area of one thirtieth of the world oceans and contains about 26.6 10^6 km^3 (Bamber and Huybrechts, 1996), equivalent to a global mean sea level rise of more than 60 m if melted completely. The annual surface accumulation is estimated to be of the order of 2200 Gt, equivalent to a 6 mm layer of water covering the entire area of the oceans. Global sea level appears to have risen during the last 100 years by about 10-25 mma (Warrick *et al.*, 1996). Although thermal expansion of the ocean, changes in surface and ground water storage and melting of glaciers and ice caps are thought to have contributed to sea level rise, a potentially major source of water for the current sea level rise is undetermined. Thus, the Antarctic ice sheet might be a major source of water for the present-day rise in sea level, but unfortunately it is not yet clear how much of the snow accumulation is returned to the sea through the calving of icebergs and the melting of ice shelves. A large positive mass balance of Antarctica and Greenland would seem unlikely, as this would have led to a substantial sea level lowering and would therefore be highly inconsistent with the estimated sea level rise (Warrick *et al.*, 1996). It is fundamental to know through an assessment of the mass balance, whether the Antarctic ice sheet today contributes towards, or instead retards, the rise in sea level, and even more, what its role will be if there is a further increase in global warming.

1.3.5.2 Estimates of present ice sheet mass balance. The ice sheet mass balance is the sum of the net accumulation across the ice sheet and the mass flux discharged at the margin (Figure 1.27). The equilibrium line is currently at or 'below' sea level around most of Antarctica so the ice loss is mainly by iceberg calving and basal melting of ice shelves, the rates of which are determined by dynamic processes involving long response times. Budd and Smith (1985) reviewed all the existing data on surface mass balance and ice velocity of the

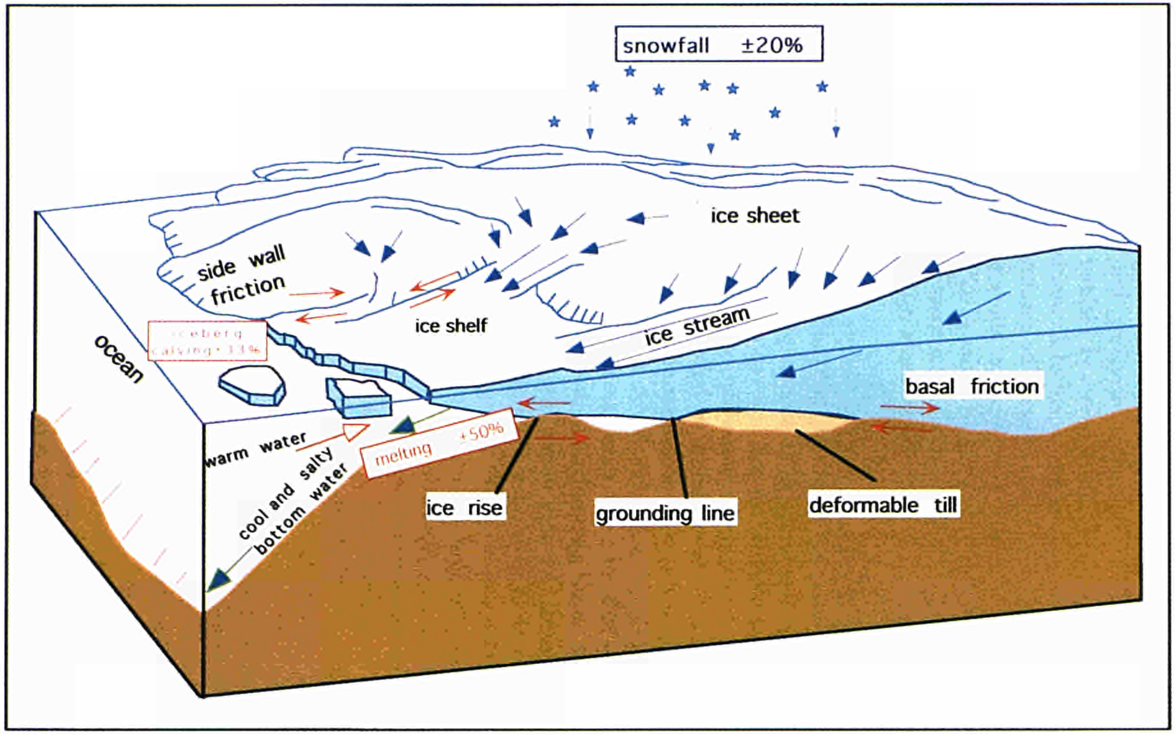


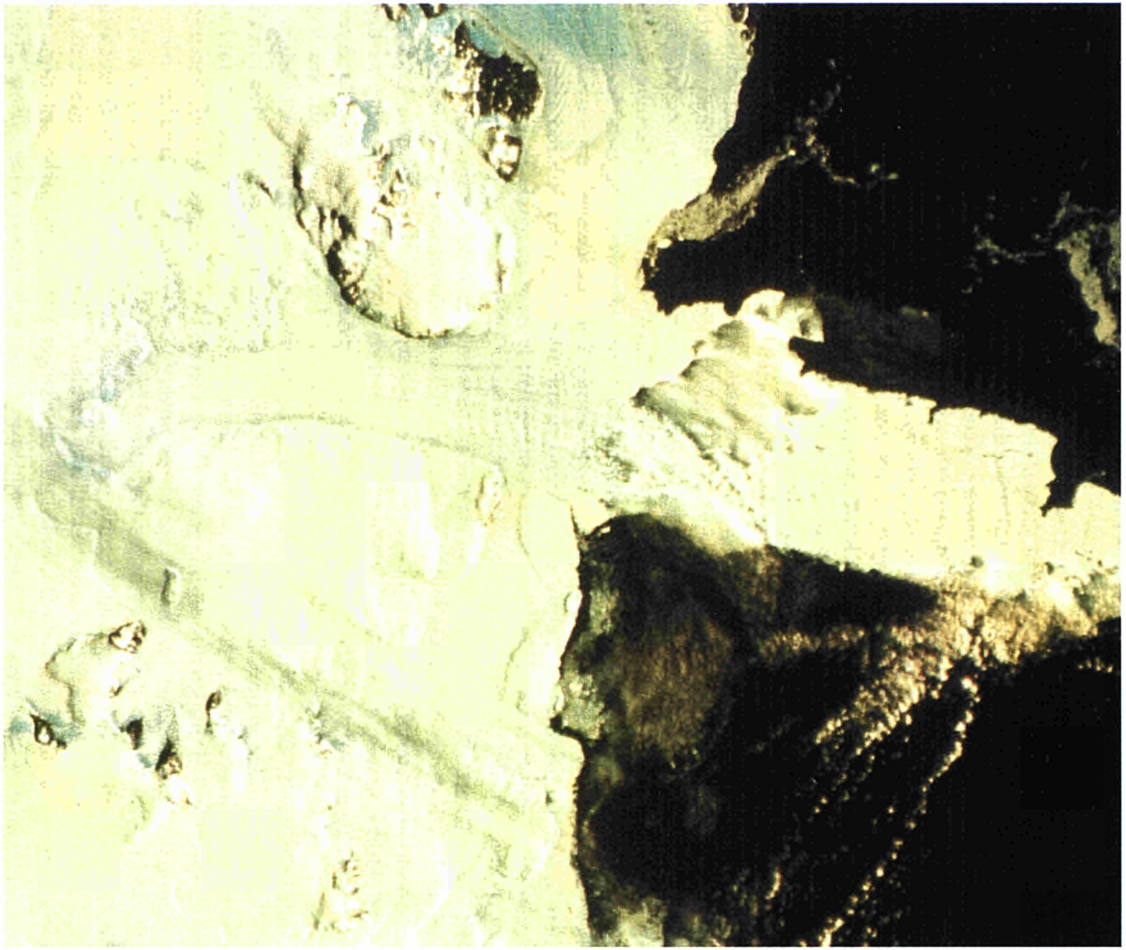
FIGURE 1.27. The principal characteristics of the ice sheet and ice shelf dynamics and estimated errors of mass balance components.

grounded ice sheet. They concluded that the total snow accumulation is nearly balanced by the outflow, although on the basis of the present data (still poor), an imbalance in the range of 0 to +20% (360 Gt/yr) would be hard to detect. Bentley and Giovinetto (1991) estimated the overall mass balance by comparison of the best available data on input in the form of snow fall with output in the form of ice flux through gates at or near the margin of the ice sheet. They concluded that there is probably a positive imbalance between +40 and +400 Gt/yr, which represents from 2% to 25% of the total yearly input. Both these studies thus suggest that grounded ice volume is increasing, but the error bars are very large. In other studies, Jacobs *et al.* (1992) and Jacobs and Helmer (1996) have calculated the present ice budget for Antarctica from measurement of accumulation minus iceberg calving, run-off and in situ melting beneath the ice shelves. The resulting negative mass balance of -681 Gt/yr differs substantially from the other estimates above but some components are subject to high temporal variability and budget uncertainties of 20%-50%. Warrick *et al.* (1996) reports that the uncertainty in the estimate of the total mass balance is at least ±25% (ca 500 Gt/yr), equivalent to a global sea level change of about ±1.4 mm/yr.

Estimates of snow accumulation of the entire continent are determined by integration of sparse measured point values and models that tie precipitation to factors such as elevation, continentality and temperature. These data present considerable problems with the quality, the inadequate spatial coverage and time-representativeness, and moreover few measurements exist of long-term changes in accumulation rate (Giovinetto *et al.*, 1989; Drewry, 1991). Nevertheless, snow accumulation is believed to be known better than other balance components. Antarctic snow accumulation values range between 2406 Gt/yr (Huybrechts and Oerlemans, 1990) and 1749 Gt/yr (Meier, 1983). Jacobs *et al.* (1992) have ascribed accuracies of ±20% to total snow accumulation. Large interdecadal changes have been reported in the snow accumulation rates across some regions of East Antarctica and the Antarctic Peninsula. Ice core studies showed variability on time scales up to a century for results from South Pole (Mosley-Thompson *et al.*, 1995), Dome C (Pourchet *et al.*, 1983) and Wilkes Land (Morgan *et al.*, 1991) suggesting a secular increase (20-30%) in annual snow accumulation for the East Antarctic ice sheet over the last 30 years. Studies of ice cores and mass balance in the Antarctic Peninsula (Peel, 1992; Raymond *et al.*, 1996) show a comparable increase over the last 200 years, with a gradual increase in accumulation rate since the 1950s. The generally rising temperatures detected in the Antarctic Peninsula correlate with increasing rates of accumulation at drill sites and with an increasing frequency of days of precipitation recorded at Faraday Station (Peel *et al.*, 1988).

Outflows across or near the grounding line have been measured in only a few drainage basins (Bentley and Giovinetto, 1991), usually some distance inland of the grounding line or near the front of an ice shelf (Figure 1.28). In the latter case, some interpretation of the regime of the ice shelf, particularly the basal melting rate, is necessary in order to estimate the net mass balance inland of the grounding line. Iceberg calving and in situ melting and freezing processes have a strong bearing on the dynamics of ice shelves, which in turn influence the positions of the grounding lines and the rate that ice flows off the continent (Hughes, 1973). Calving and basal melting, if not balanced by ice advection, decrease the dimensions of an ice shelf and its ability to retard the seaward motion of land ice (Jacobs *et al.*, 1992). Large uncertainties accompany the negative terms of the mass balance. For example, Jacobs *et al.* (1992) have estimated errors of ±33% for iceberg calving and ±50% for ice shelf melting and surface runoff..

FIGURE 1.28 Landsat Thematic Mapper images of Terra Nova Bay area showing the David Glacier with its Ice Tongue (90km long and flowing West to East) Drygalski and Nansen Ice Sheet feeds from Priestley and Reeves outlet glaciers. Most of the Antarctic Ice Sheet is drained by outlet glacier and ice streams, although only 13% of the Antarctic coastline consists of these glaciers, their drainage is about 90% of snow that falls in inland of the coastal zone.



1.3.5.3 Changes in the volume of the Antarctic Ice Sheet. In recent years, some ice fronts have dramatically receded. Major changes have occurred around the Antarctic Peninsula, as shown by break-up of several ice shelves (Vaughan and Doake, 1996; Rott *et al.*, 1996). Vaughan and Doake (1996) pointed out that ice shelf extent may well be a sensitive indicator of regional climate change. They found that northern ice shelves of the Antarctic Peninsula have retreated dramatically in the last 50 years. Meteorological records from this region show a spatially consistent warming trend since 1945, with annual mean temperatures that have risen by 2.5 °C, a far larger rise than seen elsewhere in the Southern Hemisphere (King, 1994). The limit of ice shelves apparently corresponded with 0 °C mean monthly January temperature, given by proxy in the climate records as the -5°C mean annual temperature (Vaughan and Doake, 1996). The warming observed around the Antarctic Peninsula corresponded to a southward migration of the -5° isotherm of 200 km on the west side of the Peninsula. Where the limiting isotherm has intersected the ice shelf, the shelves have disintegrated. Meltwater percolation destroys the winter's cold wave by refreezing and raises the ice shelf to the pressure melting point (Mercer 1978). Frezzotti (1997) pointed out that Hells Gate ice shelf and the McMurdo Ice Shelf in the Ross Sea area have undergone a significant retreat since the beginning of the 20th century. The retreat must be due to increased energy available for meltwater production of marine ice that progressively warmed these thin ice shelves and then increased iceberg calving. Lack of ice shelves in areas where there is substantial melting at the coast supports the idea that melt-induced changes in structural integrity prevent ice shelves from being stable.

The West Antarctic ice sheet is the world's only remaining marine ice sheet, an ice sheet anchored to bedrock below sea level and with margins that are floating. The ice shelf is important in restraining and, perhaps, stabilising discharge from the continent (Thomas, 1985). Model simulations of the West Antarctic ice sheet suggest that sporadic, perhaps chaotic, collapse of the ice sheet occurred throughout the past one million years (MacAyeal, 1992). If the entire West Antarctic ice sheet were discharged into the ocean, the sea level would rise by 5 to 6 m. Warrick *et al.* (1996) pointed out that estimating the probability of such an event is not yet possible. Bentley (1997) believes that a rapid rise in sea level in the next century or two from a West Antarctic ice sheet cause could only occur if a natural (not induced) collapse of this ice sheet is imminent, the chances of which, based on the concept of a randomly timed collapse on the average of once every 100,000 years, are on the order of 0.1%. Moreover, there is some evidence that the glaciological characteristics of the West Antarctic ice sheet can change significantly on the time scale of a few decades. Recent field studies of the West Antarctic ice sheet reveal that rates of discharge from some of the major ice streams have changed markedly, with sudden changes occurring in the last thousand years. The flow of ice stream C abruptly stopped approximately 150 years ago (Retzlaff and Bentley, 1993), while adjacent, ice stream B underwent a sudden lateral jump of ten kilometres (Bentley *et al.*, 1994) and is discharging 50% more ice than it receives in snow accumulation (Bindschandler *et al.*, 1993). Downstream of ice stream B along the Ross Ice Shelf, the Crary Ice Rise is migrating upstream and changing the regional velocity field (Bindschandler, 1993), and elsewhere, along the Walgreen Coast (West Antarctica) the flow of Thwaites Glacier is accelerating (Ferrigno *et al.*, 1993). All these complex features are large enough to cause balance changes which illustrate that the ice sheet is in a non-equilibrium state.

Undoubtedly, the volume and geographic extent of the Antarctic ice sheet have undergone major changes over geological time. During the Last Glacial Maximum (ca 20,000 years ago) the Antarctic Ice sheet was considerably larger, and retreated to near its present geographic extent about 6000 years ago (Drewry, 1979; Denton and Hughes, 1981). It is likely that the Antarctic ice sheet is still reacting to the glacial-interglacial transition between 20,000 and 10,000 years ago, and subsequent increase in the snow accumulation. Huybrechts (1990) pointed out that the Antarctic ice sheet shows a very strong response to the glacial-interglacial transition, so if there has been a significant Antarctic contribution to the recent sea level change, it is still not clear whether this is a long-term trend or a response, for example, to changing accumulation on a century time scale.

The onset of global warming could have a number of important different impacts on the Antarctic ice sheet. These include increasing basal melt of ice shelves, faster flow of grounded ice, increased surface ablation in coastal regions, and increased precipitation (Budd and Simmonds, 1991; Genthon, 1994b; Huybrechts, 1994b). Care has to be taken to separate contributions from grounded and ice shelf areas. Furthermore it is clear that the ice sheet cannot be treated as a single unit with common responses and characteristics (Fortuin and Oerlemans, 1990; Drewry and Morris, 1992). Analysis of these separate terms by ice sheet modelling indicates that the impact of increasing ice sheet flow rates on sea level does not become a dominant factor until 100-200 years after the onset of warming. For the next 100 years the most important impact on sea level from Antarctic mass balance can be expected to result from increasing precipitation minus evaporation balance over grounded ice (Budd and Simmonds, 1991; Fortuin and Oerlemans, 1992). The impact of global warming can be expected to cause a large decrease in the ice shelves but have little impact on the grounded ice over the shorter term of about 100 years (Budd *et al.*, 1994; Huybrechts, 1994b). Drewry and Morris (1992) estimated that present snow accumulation in the Antarctic Peninsula is about 25% (ca 500 Gt/yr) of that in the entire Antarctic continent. They estimated the potential contribution to sea level due to increased runoff in that part of the Peninsula where melting may occur, suggesting that for a 2° C rise in mean annual surface temperature over 40 years, ablation in this region would contribute at least 1.0 mm to sea level rise, offsetting the fall of 0.5 mm contributed by increased accumulation. Nevertheless, more comprehensive observational field data sets are required to provide better validation of the model simulations of the present regime, particularly the flow of the large ice streams, the ice thickness and the current state of mass balance. For example Bamber and Huybrechts (1996) have obtained improved data sets of ice-thickness and bed-elevation of the Antarctic ice sheet using ERS1 radar-altimeter data. The new total volume of ice is 12% smaller than the previous estimate by Drewry (1983), the most commonly cited publication by numerous investigators. The difference is equivalent to 8 m in global mean sea level and is more than the entire volume of the Greenland ice sheet. The dominant uncertainties for future conditions still rest largely on the uncertainties in climatic forcing. Fully coupled atmosphere, sea ice and ocean models, with high resolution around the Antarctic continental shelf, will be needed to derive accurately present heat and mass transfers, so that simulation of future scenarios, can be realised (Budd *et al.*, 1994).

1.3.5.4 Research objectives. Realistic projections of changes in global sea level require knowledge of the present mass balance of polar ice sheets. The vastness and environmental conditions of the ice sheet are the major obstacle in mass balance studies. The high logistic costs of working in Antarctica have prevented any one nation from attempting studies of the mass balance of the entire continent. In order to coordinate mass balance research, a Group of Specialists on Global Change and the Antarctic (Gos/GLOCHANT) was formalised in 1992 by the Scientific Committee on Antarctic Research (Weller, 1993). Science plans are being progressively established to coordinate priority research tasks, in conjunction with IGBP, SCOR, WCRP and START, and newly established programmes on Ice Sheet Mass

Balance and Sea Level (ISMASS) and International TransAntarctic Scientific Expedition (ITASE), see Figure 1.29.

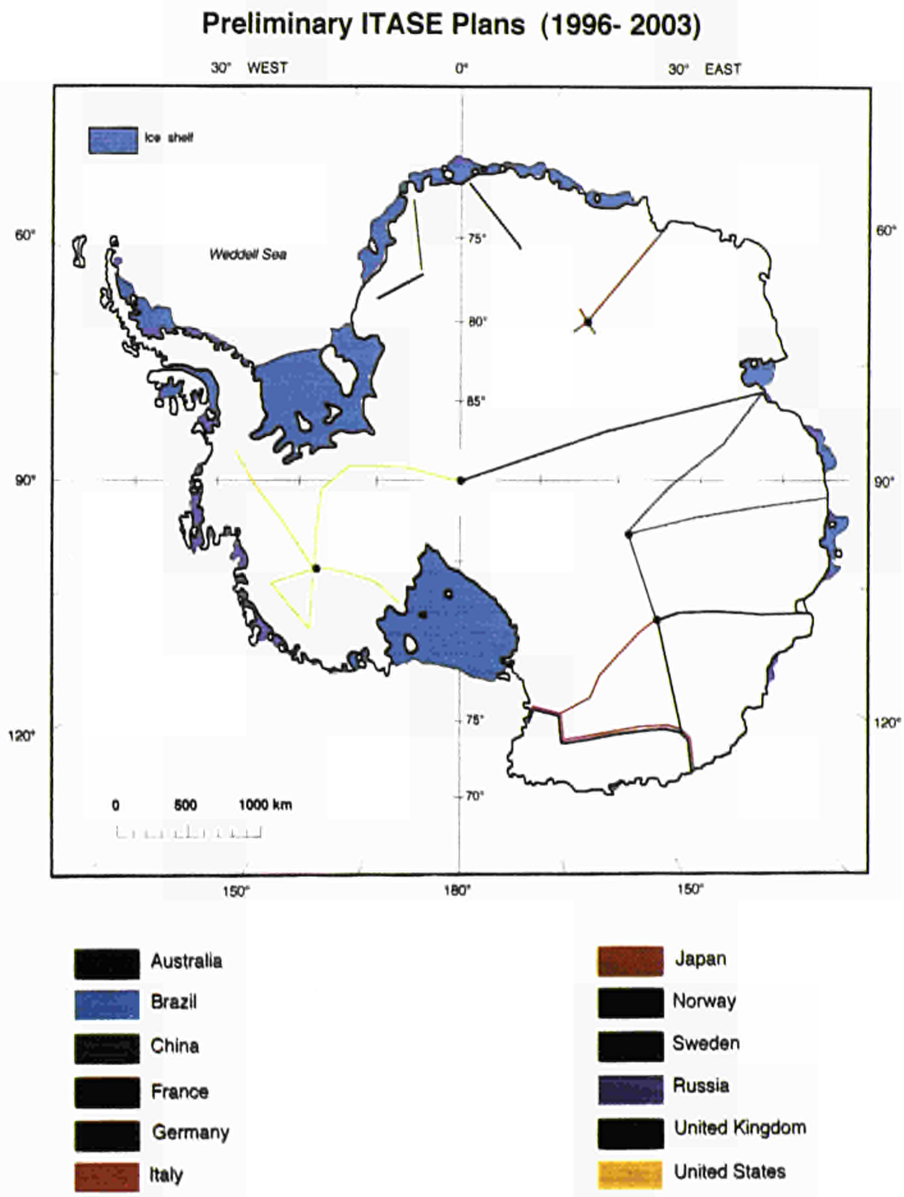


FIGURE 1.29 Map of planned and proposed National and multinational ITASE traverses for the period 1996-2003. The broad aim of ITASE is to determine the spatial variability of Antarctic climate (e.g. accumulation, air temperature, atmospheric circulation) and the environmental variability in Antarctica over the last 200 yrs.

The programme ISMASS (Goodwin, 1996) identifies the following principal objectives to encourage and promote national and international work to improve estimates of mass balance:

- conduct ice thickness measurements along grounding lines with the aim of producing continuity of measurements around the continent;
- encourage and coordinate the measurements of velocities by analysis of sequential satellite imagery around the continental periphery;
- encourage more scientists to develop the expertise needed to 'use the SAR interferometric technique and apply it to the measurement of glacial velocity in Antarctica;
- assist in the development of the ITASE programme that is a multi-National programme whose aim is to develop a high resolution interpretation and 3-D map to determine the spatial variability of climate (e.g.. accumulation, air temperature, atmospheric circulation) and environmental variability (sea ice variation, ocean productivity etc.) in Antarctica over the last 200 years;
- encourage further theoretical development and application of passive microwave radiometry of determining surface mass balance;
- promote theoretical and applied advances in calculating the snow accumulation rate in the Antarctic interior using the moisture-flux-divergence technique.

1.4 DISCUSSION AND CONCLUSIONS

Oxygen isotope stratigraphy of benthic forminifera provides a valuable record of ocean volume (rather than sea level) changes for Late Quaternary (last circa 120,000 yrs). However, the oxygen isotope record of ocean volume changes since the last glacial maximum is not particularly detailed owing to the effect of seafloor bioturbation. For this time interval, several records of submerged coral sequences from low latitudes provide valuable data on rates of ocean volume changes and of relative sea level changes for particular areas. These analyses form the basis for the proposal that "eustatic" sea level rise since the LGM was punctuated by a series of catastrophic rise events that have, in turn, been related to characteristics of regional ice sheet melting and meltwater discharge histories. The magnitude and timing of catastrophic sea level rise events considered to have taken place since the LGM has not been unequivocally demonstrated on low latitude coasts, largely due to uncertainties regarding the water depths in which specific corals grew and also to sampling and dating problems.
Similarly, attempts to constrain regional ice sheet histories are beset with problems. For some of the largest ice sheets (e.g. Greenland and Antarctica), the patterns of ice volume change since the LGM remain uncertain whilst the reconstruction of former ice sheet surface profiles (e.g. Laurentide) depends almost exclusively upon mathematical estimates. Thus the estimates of changing dimensions of ice sheets since the LGM presents a major source of error in the estimation of former ocean volume changes.

The accuracy of the models of sea level change based on models of earth rheology are constrained by the above sources of error. These studies depend on a well-constrained calibration between the calendar year, radiocarbon year and Uranium-Thorium timescales. At present considerable research is being undertaken in this regard, and it is hoped that significant progress will be made in the future once the accuracy and calibration of the respective timescales is improved, and an enhancement and enlargement of the Late Glacial (18ka-10ka) sea level data base achieved.

Estimates of the twentieth-century trend in global mean sea level from the longer tide gauge records in the PSMSL data are approximately 18 cm/century (+/- 7cm/century). This uncertainty reflects the different methods used for the removal of vertical land movements and statistical uncertainty. It does not, however, fully reflect the poor spatio-temporal distribution and varying quality of the tide gauge records. Spatial coverage of the PSMSL data set from the last century is primarily a northern hemisphere coastal one and since we do not expect long term sea level changes (whether 'trends' or 'accelerations') to be exactly the same everywhere because of changes in the ocean circulation and ocean loading, tide gauge based estimates may not precisely reflect changes in the ocean mass and volume. Nevertheless, the similarity of secular trends from widely separated locations does support a global increase in the last century. Moreover, conversely, the tide gauge rates can provide limits to estimates of ice sheet fluctuations in that time. Nevertheless, the similarity of secular trends from widely separated locations does not change a global increase in the last century. Moreover, the tidal gauge rates can provide limits to estimates of ice sheet fluctuations in this time. Recent sea level measurements with high and homogenous spatio-temporal coverage have been obtained with satellite altimetry. But since there is evidence for significant interannual variability in the global ocean water volume (Zwally and Giovinetto, 1997) there must be caution against drawing premature conclusions from short series. For the future, monitoring by coastal tide gauges and by satellite altimetry in the deep ocean could provide better estimates of the changes in ocean volume. One incentive for making reliable observation based estimates of global mean sea level change over the past century is to provide a check on model based estimates of changes in the mass and volume of the ocean resulting from climate forcing. Altimeter data sets will, however, require continuous inter-calibration with gauge data in order to provide quality control of the remotely-sensed data, and to provide historical context.

Estimates of the contribution to sea level rise from thermal expansion, glaciers and ice caps, and surface/ground water are about 8±10cm over the last century (Warrick *et al.*, 1996). For agreement between this estimate and the trend over the last century obtained from tide gauge data, the combined contribution from the Greenland and Antarctic ice sheet would have to be in the range -9 to +26 cm. This indicates that the ice sheets together are more likely to have caused a sea level rise, though the possibility of a fall cannot be excluded. The uncertainty in the recent mass balance of the Greenland and Antarctic ice sheets corresponds to a rather arbitrarily set imbalance of 25% of the mass turnover or ±18 cm/century (Warrick *et al.*, 1996). It is likely that the ice sheets are still reacting to the last glacial -interglacial transition. In the case of East Antarctica and the Antarctic Peninsula there is, in addition, evidence of increased accumulation over the last 30 years or so. Sorting out the background trend from changes on time scales of a century or less is a challenge. Science plans are being progressively established to co-ordinate priority international research tasks.

The uncertainty in the present mass balance of the ice sheets (and also to a lesser extent of the glaciers and ice caps) contributes to the very large uncertainty in projections of ocean volume change for the next century. Important uncertainties also arise from uncertainty in future emissions, in the climate sensitivity and in future regional climate change as well as in the modelling of anthropogenic changes in thermal expansion and land ice volumes. Projections over the next century for the 1-D model, including many of these uncertainties, range from

13cm to 110cm with a mid estimate of 55 cm (IS95a, constant 1990 aerosol case). Results for the 2-D model for a range of emission scenarios (IS95c - IS95e), but with a climate sensitivity of 2.2 °C and with best guess ice model parameters only, range from 17cm to 49cm, with a mid estimate of 34cm (IS95a, constant 1990 aerosol case). Ocean volume changes expressed as sea level rise projections for the CO_2 stabilisation scenarios show no sign of levelling off by the year 2300, when a climate sensitivity of 2.5 °C or greater is assumed (Wigley, 1995). While these longer projections must become increasingly tentative, they raise the question as to the sea level rise commitment inherent in the stabilisation scenarios.

It is noteworthy that, although sea level rise projections are generally given with a range of uncertainty, this range does not generally represent a particular probability range in a formal sense. Clearly when uncertainties for a large number of parameters are concatenated the resulting probability for the extreme low and high estimates would have a very low probability of being realised. Thus in the 1-D model case, for example, the parameters were carefully selected with the intention of giving a range which represents approximately the 90% probability range.

Changes in ocean volume are only one of the factors which lead to regional or local changes in sea level. Other factors are the subject of the next section.

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2

REGIONAL AND LOCAL SEA LEVEL VARIATIONS

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2.1 INTRODUCTION

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A "global sea level curve" and particularly a "global sea level rise" plays a prominent role in the current climate-change debate. Particularly for the assessment of future impacts of climate change and the planning of actions to mitigate these, a knowledge of future relative sea level changes is of high social and economic value. Consequently, the possibility of a significant sea level rise over the next 50 to 100 years has invoked considerable public interest. Furthermore, it has been established that contemporary coastal process functioning and morphological development result from a linkage to a number of factors controlling sea level at a range of time and spatial scales, the parameterisations of which are of fundamental importance in understanding sea level as a control of the coastal zone.

Attempts have been made to identify a "climate signal" in global sea level, which can be expected for two reasons: (1) a change in the heat content of the water of the oceans leads to a volume change (this effect has entered the discussion under the expression "thermal expansion"); and (2) mass exchange between ocean and the cryosphere. Therefore, a knowledge of the global ocean water mass and volume (GOMV) as function of time constitutes a crucial constraint for the reconstruction of past climates, as well as the validation of models used to predict future changes in the GOMV. However, due to the fact that the Earth is a deformable, body with gravity, there is no simple relation between sea level observed at any point on the Earth's surface and the GOMV. Up to now, establishing past variations in the GOMV has been hampered by the complexity of this relation, and the available global curves and trends are still insufficient.

"Sea level curves" for the last 20,000 years have been established from isotopic and geological data which have been used as proxies for ocean mass and volume changes (see e.g. Pirazzoli, 1996). Moreover, these curves have been used as input for geophysical sea level models and have been iteratively improved (see e.g. Peltier, 1994). Nevertheless, the current knowledge of past volume and, particularly, mass changes of the oceans due to the disintegration of the ice sheets of the last ice age is still unsatisfactory and the large uncertainties in the GOMV function over time constitute a serious limitation for our capabilities to predict future changes in this function. The importance of sea level changes (rates and patterns) has also been recognised as a primary forcing control in coastal functioning, understanding of which is critical in the development of coastal zone management policies. Much more information at the level of local to regional scale effects of such sea level control is required. Of particular importance here, are improvements in the linkage and modelling of the long-term factors (100 to 1000 years) in sea level movements, which are time-lagged in their influence upon coasts, with contemporary sea surface controls.

Sea level curves for the last 100 to 150 years, as well as global sea level trends have been determined in several studies (for a review, see e.g. Warrick *et al.*, 1996). However, all of them used basically the same global data set of monthly (or annual) mean sea level records obtained from coastal tide gauges (see sections 2.2.2 and 2.3.1). Despite the common data base, the resulting global sea level trends display a considerable scatter of between 1 and 2.5 mm/yr indicating the degree of freedom in data preparation, selection, and corrections inherent both in the data set and the complex relationship between relative coastal sea level at a single tide gauge and the GOMV. Relative sea level at any given tide gauge may be affected by many causes operating on different spatial and temporal scales which separate into two broad classes, namely those changing the sea level (with respect to a fixed reference surface) and those resulting in vertical land movements. Only a rigorous analysis of the global tide gauge data set making use of all available information concerning local effects at the tide gauge location can determine reliable uncertainty bounds for the global signal. Up to now, no such rigorous but urgently required analysis is available. The uncertainties given for the global trends determined in the various studies are of the order of less than 1 mm/yr, but these uncertainties are pure statistical quantities. The uncertainties due to the complexity of the physical problem may in fact be much larger than the numbers given.

The presently available results from the TOPEX/POSEIDON (TP) mission reveal a spatial variability in the local sea level trends over the period of observation (five years) of the order of ± 50 mm/yr (see section 2.4). However, evidence from longer tide-gauge records indicates that this pattern is due to decadal sea level variability. This variability is of the same order of magnitude as the global mean sea level trend determined from the three years of TP observations. A detailed study of the spatial variability at decadal time scales in relation to other parameters such as the sea temperature and the wind field is required to improve our understanding of these variations and their relation to the globally averaged sea level trend.

There is a kind of trade-off in the mutual uncertainties in both, determination of changes in GOMV and determination of the major contributing factors, i.e. the mass and heat balance of the ocean.

The ocean is the major reservoir of the global hydrological cycle. Among other parameters such as its geometrical shape or its dynamic state, this reservoir is characterised by its mass and volume. Consequently, GOMV may be affected by all changes in the other reservoirs of this cycle such as groundwater, soil moisture, humidity of the air, terrestrial surface water, ice sheets, glaciers, or permafrost. Presently, human interference such as deforestation, groundwater extraction, irrigation, river basin developments or reduced infiltration due to road infrastructure and urban development are at a level where the global hydrological cycle is significantly affected. Studies of these anthropogenic influences in particular have demonstrated that some of them may significantly contribute to GOMV changes. However, there are considerable differences between estimates obtained in the various studies basically because they were based on insufficient data. Therefore, more thorough studies of the fluxes and reservoir changes in the hydrological cycle based on the broadest possible database are necessary to clarify the contribution of the various processes to GOMV changes.

Many parameters characterising the hydrological cycle (such as ground moisture, effects of infiltration on groundwater renewal, percolation of groundwater, subsurface discharge of groundwater into the ocean) are known within large uncertainty limits. The uncertainties are basically due to spatially and temporally insufficient observations of the transport of water within nearly all components of the hydrological cycle. To a large extent, this lack of observations is caused by the absence of technologies that allow sufficient monitoring of the relevant parameters within given economic constraints.

Considering the importance of the hydrological cycle for the functioning of the biosphere as well as most near-surface processes of the Earth system, and its role as a major constituent of the climate engine, the need for innovation in techniques applicable to the monitoring of water-mass movements cannot be over-emphasised. There is a clear need for novel approaches in this field including, for example, gravity missions designed to detect relatively small mass movements in the Earth system.

A major contribution to the ocean mass changes in the past clearly resulted from changes in the cryosphere, particularly in the large ice sheets. However, the uncertainties in our knowledge of the current mass changes in the cryosphere are too large to constrain even the sign of the contribution (see Warrick *et al.*, 1996). A reliable prediction of the future GOMV changes or trends will only be possible if the uncertainties in our knowledge of the current mass changes in the cryosphere can be narrowed. Moreover, the long (centennial) time scales dominating the response of ice sheets to climate variations pose a serious challenge for dynamic models of ice sheets, and considerable improvements of the models are required particularly with respect to their ability to model the long-term behaviour of ice sheets. Studies based on General Circulation Models (GCM) and observational evidence suggest that climate change as well as the response of the ice sheets to this forcing are spatially highly variable. Currently, mass balance calculations for ice sheets are based on forcing functions averaged over large regions such as, for example, Greenland. It is widely accepted that this large-scale approach may be inappropriate and more detailed calculations with high-resolution forcing functions are urgently needed.

Glaciers exhibit a broad spectrum in size, and the response of glaciers to climate change can be expected to be also dependent on the size of the glaciers. Up to now, most studies have concentrated on smaller glaciers. To cover the broad spectrum of glaciers, special efforts should be made to include large glaciers in future studies.

Monitoring the changes of the surfaces of the large ice sheets still poses considerable problems to the remote sensing methods currently in use including satellite altimetry. Considerable technological innovations comprising novel approaches are required to reduce the uncertainties in the observations related to mass changes of the ice sheets and to provide the technological means for an economically feasible monitoring of this highly relevant parameter.

Besides mass exchange with the cryosphere, the volume increase due to thermal expansion of the ocean is considered as a major contributor to future GOMV changes. However, the uncertainties associated with the estimates of the effect of thermal expansion on GOMV over the last 100 years are still large. The uncertainties mainly result from a lack of sufficient observations of the present state and changes in the three-dimensional temperature and salinity in the ocean. The models used for estimating thermal expansion range from three-dimensional atmosphere/ocean general circulation models (A/OGCM) to simple one-dimensional models. Improvements in the parameterisation of heat penetration into the ocean is needed in all these models with the improvements being based on observations and inter-model comparison.

Observations of sea level variations on different time scales are highly relevant. On the one hand, they allow for the analysis and description of sea level variability which is a prerequisite to a better understanding of the causes behind these variations. On the other hand, they constitute crucial constraints for models related to sea level, such as hydrodynamical models or coupled atmosphere-ocean circulation models.

Tide gauges (see below) have been operated at coastal sites since the beginning of the nineteenth century in an ever-increasing number. Today, approximately 2,000 tide gauges are operating globally. The data set of monthly mean values computed from the tide gauge records and collected at the Permanent Service for Mean Sea Level (PSMSL, see http://www.nbi.ac.uk/psmsl/psmsl.info.html/) constitutes one of the most valuable climatological data sets. In particular, the long records with spans of several decades are a unique source of knowledge for the interannual to centennial sea level variability, which is directly related to climate variability. It is of the utmost importance not to let this data set be interrupted. In order to maintain or even improve the spatial coverage of coastal tide gauges, it is also required to continue to operate tide gauges with shorter records or even set up new tide gauges in areas which up to now have no records. The Global Sea Level Observing System (GLOSS) has put forward an implementation plan which supports this by giving exact locations for a global tide gauge network fulfilling both the long-term and the spatial aspect.

Satellite altimetry has proved to be a most valuable source of information on intra- to inter-seasonal sea level variability with a near global coverage. Current and future missions strongly rely on tide gauge calibration. For this purpose, suitable tide gauges have to be co-located with space-geodetic equipment such as GPS or DORIS receivers, which provide a high-accuracy control of the tide gauge benchmark movements. Here again the GLOSS implementation plan includes a list of suitable stations as well as recommendations for the observational strategies. GLOSS also encourages regional densification of the tide gauge network and has accepted proposals for EUROGLOSS and MEDGLOSS networks (including GPS, absolute gravity, etc.). It is strongly recommended to support the implementation of GLOSS both on national and international level.

TOPEX/POSEIDON has revolutionised our knowledge of the spatial variability of sea level on intra- to inter-seasonal time scales and improved our understanding of the forces producing this variability. At decadal and longer time scales, coupled atmosphere-ocean phenomena contribute dominantly to the climate variability. Therefore, extending these sea level observations to decadal time scales would provide an invaluable key to a better understanding and eventually to a long-term prediction of climate variability at inter-annual to decadal time scales. Thus, a continuous programme of TOPEX-class altimetry of the sea surface will provide observations crucial in understanding the forcing of climate variability at inter-annual and decadal time scales.

Tide gauges measure sea level relative to a benchmark on land. Separation of crustal movements and the "oceanic" contribution to the relative sea level changes (RSL) is only possible to the extent and accuracy to which the vertical movement of a benchmark is known. Vertical crustal movements of coastal sectors may occur over a broad range of spatial and temporal scales (Aubrey and Emery, 1993) and may result from several natural processes (glacio- and hydro-isostasy, sediment loading and compaction, lithospheric cooling, neo- and volcano-tectonics), locally enhanced by anthropogenic effects (mainly extraction of groundwater, oil or gas). All these processes form a significant component of the short-wavelength (i.e. < 100 km, according to Gornitz, 1993), secular to decadal RSL variability, which affects both regional and local RSL changes with rates of the same order, or even in excess (up to several mm/yr), of the long-term trend resulting from changes in GOMV. Their contribution may become particularly evident during periods of relative small changes in GOMV, such as the last 6 kyrs (Pirazzoli, 1996).

In addition to possible non-linear trends, crustal movements in most cases show short-scale spatial variability; this should be considered in terms of geographical bias and undersampling of geological variability as, for example, resulting from the present non-uniform distribution of tide-gauge stations along coastal margins (Emery and Aubrey, 1991; Gornitz, 1993). The knowledge of the local geology and the collection of geologic evidences of RSL in the past (last 20,000 yrs or before) should be used to evaluate the reliability of regionally averaged RSL curves and to verify how far the trend of local RSL change could be extended laterally along the coast from a tide gauge.

Space-geodetic observations at the tide gauges provide the best available means to observe the benchmark movements with sufficient accuracy. Following the recommendation of respective international working groups, tide gauges to be used for observations of sea level with respect to a fixed reference surface have to be co-located either permanently or episodically with GPS receivers and absolute gravity measurements should also be made wherever possible.

2.2 OBSERVATIONAL STRATEGIES: THE TECHNIQUES

2.2.1 Geologic markers of the past

C. Romagnoli

Observations on past trends in the relative sea level are based upon geomorphological, stratigraphical and historical-archaeological records, which may witness past sea levels as well as neotectonic activity along the coasts. Late Pleistocene and Holocene palaeo-sea level indicators are used to determine the long-term component of crustal motions, which have to be filtered out from RSL trends. Rates of sea level rise obtained during the last deglaciation (since the last 18,000 yrs) represent a fundamental basis for comparison with the historical and present-day ones and for verifying the contribution, over the last 100 yrs, of climate-related or anthropogenic factors. Nevertheless, studies on sea level fluctuations in the past clearly indicate that historical records of RSL are not comprehensive of all patterns of short-term sea level variability. For instance, very rapid pulses in the sea level rise have been recognized around14,000 and 11,300 yrs ago, when sea level rose abruptly at mean rates of 40-50 and 20-30 mm/yr respectively in correspondence of periods of accelerated melting (Bard *et al.*, 1996); rates of sea level rise of similar order have been reported also for the southern North Sea during the Eemian Stage (Zones E3b-E4a; Roe, 1997).

Markers of former sea levels along coastal sectors are represented by various kinds of indicators, whose effectiveness in determining the position of RSL in space and time mainly depends on how much they can be assumed to represent relatively narrow limits of depth and contain material datable by means of radiochronological techniques.

Erosional forms along the coasts (such as tidal notches or wave-cut platforms) and bioconstructional/erosional activity (organic growth, grazing or borings by several organisms) represent the most common witnesses of past sea levels, together with archaeological structures lying on the coast. The most accurate RSL indicators are in situ intertidal deposits or upper sublittoral bioconstructions preserved in growth position, characterized by the narrowness of their vertical zonations and development (less than (5 cm), which are independent of exposure, lithology and steepness of the coast, currents and surf energy (Pirazzoli, 1996; Kelletat, 1997). Vermetids or calcareous algae (such as Dendropoma or Litophyllum Lichenoides) have been used, along Mediterranean coasts, to reconstruct former sea level with an accuracy of ± 0.2 m or even ± 0.1 m.

An accuracy of the same order (0.05-0.30 m) has been reported from the study of foraminiferal assemblages from macrotidal paleomarsh peat deposits (Scott and Medioli, in Pirazzoli, 1996; Gehrels *et al.*, 1997).

For the last 2000-5000 yrs the most widespread data type, especially in areas such as the Mediterranean, is represented by coastal archaeological sites, which can provide a fairly accurate assessment of the vertical rate of crustal movement averaged over 1000-2000 yrs (although this average can severely underestimate possible short-term movements; Flemming, 1992). This kind of evidence, if supported by accurate observation and instrumental methods, may be an excellent sea level indicator and may result in an accuracy of the order of ± 1 m to as little as ± 0.2-0.3 m.

Besides these discontinuous types of evidence, climate-related proxy data, such as oxygen isotope variations in pelagic sedimentary successions, represent a continuous and high-resolution record of past sea level fluctuations (although biased by variations in water temperature and oceanic circulation). Isotopic data appears to match sea level highstands derived, for the last 140,000 yrs, from raised coral terraces (corrected for tectonic uplift and water depth of growth, Bloom and Yonekura, 1985; Radtke and Grun, 1990; Pirazzoli *et al.*, 1993) and this agreement provides more precise chronological constraints (Chappell *et al.*, 1996).

It has been estimated from isotopic curve that during the Quaternary the sea level might have remained, for 75 per cent of the time, lower than -20 m and for 50 per cent of the time, lower than

-40 to -50 m (Chappel and Shackleton, 1986; Williams *et al.*, 1988). This means that most traces of ancient sea levels, unless tectonically uplifted, should be presently submerged or buried under sediments (Pirazzoli, 1987); thus, evidences of the sea level position during the last glacial maximum (about 120 m below the present sea level) and during the subsequent transgression, which led the sea level to its present position, should be investigated from underwater surveys.

Recently, the radiometric dating of speleothems sampled in submerged caves has been usefully applied as a mean for reconstructing past changes in the RSL (Richards *et al.*, 1994; Antonioli and Oliverio, 1996). Submerged speleothems are good indicators of the absence of the sea level during their formation (subaerial growth) and of the time of marine submergence during RSL rise (stop of internal growth and beginning of overgrowth from marine encrusting organisms).

In recent years, due to the development of the seismo-stratigraphic approach (Posamentier and Vail, 1988; Posamentier *et al.*, 1988), the observation of sea level signatures on clastic depositional sequences of continental margins through high-resolution seismic investigations has greatly improved, revealing a potential for resolving also short-term fluctuations of the RSL during the Late Quaternary. On the other hand, a main problem remains the investigation of buried marine sedimentary successions through direct sampling, which needs complex instrumentation and expensive drilling operations. High-frequency depositional sequences may yield a significant registration of sea level tendencies and their balance with sediment supply and regional or local vertical movements along coastal margins. Comparative studies on mid-latitude shelves (Chiocci *et al.*, 1997, for instance) clearly reflected the effects, on the stratigraphic architecture of the margins, of Middle-Late Pleistocene sea level fluctuations, highlighting their high-frequency (about 20 to 100 kyrs) and high-amplitude (magnitudes of the order of 100 m) cyclicity and their marked asymmetry (relatively slower and punctuated sea level fall stages in comparison with short-lived rise and highstand stages).

At present, the possibilities of identification of past sea level fluctuations may depend on the duration and the oscillatory character of the event, the rate of long-term sea level change before and after the oscillation, the time scale investigated, and especially the resolution of the sea level indicators and dating method used (Pirazzoli, 1996). Moreover, as pointed out by Plag *et al.* (1996), there are still several open questions related to the observational evidence used to construct former RSL and to their interpretations. It is suggested, for example, that to focus on the palaeoenvironmental conditions in which indicators of former sea level or water depth formed; this approach requires a much higher integration of the wide range of disciplines involved and an improvement of the different techniques used in sea level studies.

2.2.2 Tide gauges

Trevor Baker

The network of coastal tide gauges around Europe provides the basic data sets for determining and understanding past and future sea level changes around the coasts, on time scales from hours to a century or more. The strategies for using satellite-altimetry and the geodetic techniques, such as GPS and absolute gravity, have to take into account the distribution and quality of the available sea level information from these coastal tide gauges. These gauges provide the basic information needed for impacts of sea levels on coastlines, the RSL trends needed for coastal flood protection and the frequency and magnitudes of storm surge events. In this section we review the main types of tide gauges used for coastal sea level measurements and discuss their limitations and possible improvements.

The most common type of tide gauge still in use is the float operating in a stilling well (Pugh, 1987; IOC GLOSS Implementation Plan, 1997). The first self recording tide gauge operated on this principle and was installed at Sheerness on the Thames estuary in 1832. The well is normally attached to a vertical structure such as a harbour wall, and is connected to the sea via a narrow cone orifice or pipe. The orifice area is typically 0.01 of the cross sectional area of the well and this serves to damp out the higher frequency disturbances due to waves. The vertical motion of the float is transmitted via a wire or tape to a pulley and counterweight system. Nowadays, the data are normally recorded electronically, rather than on a chart. Due to the widespread use of these gauges, there have been many papers written on the disadvantages of stilling well-float systems. These papers emphasise the non-linear response of the stilling well, problems due to fouling of the inlet, draw-down due to waves and currents and errors due to density differences in the well. The mechanical nature of these gauges means that their operation is more labour intensive than more recently developed tide gauges. For properly maintained gauges, careful attention is given to weekly checks on the datum (i.e. the height of the water relative to the tide gauge bench-mark, TGBM) using a steel tape with electrical detection of the contact with the water in the stilling well. Ideally this should be done through a complete tidal cycle. At the same time, the performance of the stilling well should also be checked by comparisons of the tide gauge readings against a visual tide staff outside the stilling well. In calm conditions, this can be done to an accuracy of 2 cm.

Despite the many disadvantages of float gauges, it must be remembered that the best long time series of mean sea levels (typically of order 100 years) that are available in Europe are from carefully maintained float gauges in stilling wells (e.g. from the Netherlands and Scandinavia). The necessity of regular attendance for these mechanical gauges has meant that the datums are frequently checked. In contrast, the modern types of tide gauge usually have their data transmitted to an operational centre by telephone or satellite. This means that they are visited less regularly and therefore the datum of a gauge is checked only very occasionally.

There are a large number of automatic tide gauges now available from various manufacturers. Most use either the time of flight for an acoustic wave to travel from the transducer to the sea surface and back again or the pressure at a fixed point below the sea surface. One advantage of the latter is that a fixed vertical structure is not essential. Stepped sensor techniques and electromagnetic waves have also been used for sea level measurements. However, it is important to recognise that most tide gauges are installed for harbour navigation purposes, rather for scientific purposes. Most of the available transducers have no problems with resolution, but only for those gauges specifically designed for scientific objectives has sufficient attention been given to the accuracy of the datum. The papers from a recent IAPSO-IOC Workshop on Sea Level Measurements (IOC, 1992) show, from intercomparison tests between various types of tide gauge, that 15 mm differences in mean sea levels are quite common. Whilst this is smaller than the real inter-annual and decadal variabilities of mean sea levels, which are typically 50 mm, it is still larger than is desirable.

With automatic transmission of sea level data, methods also have to be found which automatically check the datum. Automatic datum switches have been used, which detect the time that sea level passes a certain point. However, these are affected by waves and so far they only have accuracies of a few centimetres. Two tide gauge designs are available where particular attention has been given to automatically maintaining a stable datum. The first uses the Aquatrak acoustic sensor. This system is used at over 100 NGWLMS (Next Generation Water Level Measurement System) tide gauges in the USA operated by NOAA (Gill *et al.*, 1992) and also at 14 sites around the Australian coast and 11 islands in the South Pacific operated by the National Tidal Facility (NTF), Australia (Lennon *et al.*, 1992). The second type uses improved sub-surface pressure gauges developed and installed around the UK and in the South Atlantic by the Proudman Oceanographic Laboratory (POL), UK (Woodworth *et al.*, 1996).

The Aquatrak water level sensor sends an acoustic pulse down a 1.3 cm diameter PVC tube and measures the travel time for reflected signals from the water surface and also from a calibration hole at a known distance (1.2 m) from the transducer. The sensor and tube sit inside a 15 cm diameter protective (not stilling) well. 1 second sampling and averaging 181 samples effectively filters out the wave energy and the data are logged every 6 minutes and transmitted every 3 hours via satellite or telephone lines. The most important effect on acoustic gauges is the increase of the velocity of sound in air of 0.18% per degree centigrade. Since the distance to the calibration reference point is accurately known, this distance and the travel time can be used to correct for this temperature dependence. However, temperature gradients in the tube, particularly for the lengths involved with large tidal ranges, can still have an important effect on the calibration and datum. Therefore two thermistors are also used to determine the gradient and these can be used to provide any necessary datum corrections (Porter and Shih, 1996). As with any new tide gauge, it is important to have overlapping data with the traditional float gauge at each site, in order to assess more fully any systematic errors in the new system. In the USA the overlap will last at least 10 years at ten key sites.

The system developed by POL uses Paroscientific Digiquartz sensors to detect the pressure at a fixed point below the sea surface. Two types of system are used. In the pneumatic bubbler gauge (Pugh, 1987) compressed air is passed down a tube until air bubbles out of a hole in the side of a specially designed underwater pressure point. A differential Digiquartz pressure transducer in the tide gauge hut measures the difference between the underwater pressure and atmospheric pressure. This difference is then related to the depth of water by the hydrostatic equation. In the second type of tide gauge an absolute Digiquartz pressure transducer measures directly at a fixed point in the water and a second absolute Digiquartz measures atmospheric pressure in the tide gauge hut. For pressure gauges the density of the sea water must be known and particular care must be taken if the sea level measurements are made in an estuary where the density is varying. The absolute Digiquartz has also been used extensively in bottom pressure recorders (e.g. Spencer *et al.*, 1993) which measure sea level variations off-shore on continental shelves or in the deep oceans in depths of over 4000 metres. Although these off-shore gauges have no reference datum, they are nevertheless extremely useful for comparisons of off-shore sea level variations with coastal or island tide gauges and with satellite altimetry measurements and for validation of hydrodynamic models.

Although pressure transducers are widely used for coastal sea level measurements, it has proved to be difficult to maintain a datum to better than 10 to 20 mm, due to problems of uncalibrated offsets and long term drifts in the transducers. The automatic datum system developed by POL uses a further pressure transducer to measure the pressure at a point at approximately mean sea level (Woodworth *et al.*, 1996). This pressure point can be readily levelled to the TGBM and the transducer, when corrected for atmospheric pressure variations, provides a flat portion for a few hours around every low tide. Since it is at a clearly defined height, this can be used to continually provide a datum check accurate to a few millimetres. Because the pressure point is near mean sea level, it also automatically corrects the mean sea level data for any variations in sea-water density. This method could also be developed to provide automatic datum checks for float and acoustic gauges.

Whilst the above acoustic and sub-surface pressure gauges can be used to provide sea level measurements accurately related to the TGBM, this bench-mark and an array of 3 to 6 auxiliary bench-marks must also be considered to be an integral part of the tide gauge system. Annual connections by first order spirit levelling to 3 to 6 bench-marks, within 1 to 2 km of the tide gauge, provide an essential check on the stability of the tide gauge installation and the harbour area. These auxiliary bench-marks also provide important back ups for reinstalling the TGBM, if it is destroyed e.g. during harbour reconstruction.

2.2.3 Vertical crustal movements

Iginio Marson & Susanna Zerbini

The study of sea level changes involves a geometrical study, the assessment of the present sea level with respect to a reference frame, and a dynamic study, which deals with mass changes in the complete hydrologic circuit. This last is rather difficult to assess because it involves the cryopshere as well as groundwater exploitation, irrigation, reservoir construction, changes in the river's flow and so on. The assessment of the present sea level still requires the use of coastal tide gauges. Tide gauges measure local sea level relative to a benchmark on land. Thus, vertical crustal movement is one of the factors affecting the local RSL signal. Tectonic movements, post-glacial rebound, sedimentation, groundwater or oil extraction, all may result in vertical crustal movements of regional or even local scale. It is worth noting that changes in the surface mass distribution in both hydrosphere and cryosphere induce also a viscoelastic deformation of the Earth which affects the global geoid and consequently the sea level. Crustal deformation processes might therefore bias regional and local studies of sea level changes. It is then mandatory to properly monitor to a high degree of accuracy vertical and horizontal movements of the earth's crust. This can be efficiently accomplished by the modern tools provided by space geodesy (SLR, VLBI and GPS). In particular, nowadays GPS is the technique which is commonly adopted to measure, either by means of continuous or episodic observations, vertical crustal movements at the tide gauge stations. Independently from space geodetic techniques, an alternative approach to monitoring site vertical movements is provided by the measurements of absolute gravity at tide gauge benchmarks. Being sensitive to variations of the vertical component, gravity constitutes a separate and independent check on the vertical crustal movements.

Even more important, temporal variations of the gravity field contribute to the assessment of models of crustal deformation and to the comprehension of the processes involved. This research topic requires long-term projects, of the order of decades. This because the geodynamic processes which might induce crustal deformation and variations of the gravity field are slow, and because the related signals are small, so that only after a certain time one can detect a signal larger than the significance threshold. This means that observational projects should initiate now and shall be running for decades so that the data will be available when necessary. It is therefore mandatory to set up the projects in the most accurate way as possible and to take all the precautions to avoid doubts or ambiguities.

2.2.3.1 GPS

Susanna Zerbini

Many text books and journal articles are available which describe the Global Positioning System (GPS) technique and its applications in quite some detail (see, for example, Ackroyd and Lorimer, 1990; Bock and Leppard (eds.), 1990; Seeber, 1993; Beutler *et al.* (eds.), 1996; Kleusberg and Teunissen (eds.), 1996; Hofmann-Wellenhof *et al.*, 1997). Therefore the following paragraphs will mainly highlight the principal characteristics of this technique, the achievements and its present limitations.

Nowadays GPS can be included among the mature space techniques such as Satellite Laser Ranging (SLR) and Very Long Baseline Interferometry (VLBI). In the early nineties this space technique has started to contribute significantly to regional and global geodynamics studies, with the major advantage in the case of SLR and VLBI that the ground segment (antenna/receiver) is available at low cost compared to those of the other techniques and it is small and easy to transport. For these reasons GPS has been widely adopted to build local and regional networks densifying the large-scale global networks of SLR and VLBI stations. This technique provides time, position and rates accurately, inexpensively and anywhere on the Earth's surface (Remondi, 1991). Another major advantage is that GPS provides continuous positioning. At present, mean daily solutions can be derived well below the centimetre level in all three components. Time series of daily solutions can then be analysed to provide significant information on short-time co-ordinate variations and, in particular, seasonal or sub-seasonal fluctuations in the height component.

The basic observation principle of GPS can be considered the determination of the pseudoranges. They can be derived from measurements of time or phase differences obtained by comparing the signal emitted by the satellite and a replica generated by the receiver. GPS is a one-way system and it utilises two clocks, the satellite and receiver clocks, which start off at different times, since it cannot be assumed that they are perfectly synchronised. This difference, which is the same for all the measured ranges, can be calculated. These biased ranges are known as pseudoranges.

GPS, being primarily a positioning system, provides high-accuracy kinematic data. The high-accuracy results are obtained through the relative positioning approach in which at least two or more GPS systems observe the same four satellites simultaneously, and the vector between two (or more) instruments is then determined. The four pseudorange measurements are necessary to estimate the antenna co-ordinates and the clocks offset. The differential technique is adopted because it allows the removal or minimisation of correlated systematic errors between receivers.

The pseudo ranges derived from carrier phase observations, rather than those from code measurements, are used to derive millimetre-accuracy vector measurements. In fact, the phase of the carrier can be measured to better than 0.01 cycles, corresponding to millimetre precision.

GPS observation biases include those associated with the signal propagation. The propagation of the GPS signal is affected by the atmosphere in different ways. The ionosphere, by containing charged particles, is a dispersive medium for microwaves. Therefore the GPS signal is influenced by the non-linear dispersion characteristics of the ionosphere. Since the effect is frequency-dependent, the measurements performed on both frequency bands L1 and L2 allow the estimation of the ionospheric effect. The use of dual-frequency receivers is necessary for high-precision geodesy, as the dual-frequency correction removes most of the ionospheric effect from both the code and carrier phase measurements.

The neutral atmosphere, which includes the troposphere, is basically frequency independent over the radio-bands of the entire electromagnetic spectrum. The molecules and neutral atoms in the troposphere influence the propagation of electromagnetic signals. The refraction in the neutral atmosphere can be separated into two components: the dry and the wet one. While the hydrostatic component of the vertical delay may be well estimated from accurate surface pressure data, the wet tropospheric delay is difficult to predict since it depends upon the atmospheric conditions along the signal propagation path, and these may not be well correlated with surface conditions. Also the water vapour in the troposphere has a rather non homogeneous distribution in time and space. The tropospheric delay has to be estimated; rather sophisticated models are available, though improvements are still necessary (by including the effect of azimuthal variations in the mapping function) and direct measurements of the water vapour content are also carried out by means of Water Vapour Radiometers (WVR) (Zerbini *et al.*, 1996).

On the other hand, being sensitive to the atmosphere, the GPS observations offer the challenging possibility of studying different properties of the atmosphere. This was not possible earlier using other space techniques such as, for example, VLBI, because of the required density in space and time of the observations. The development of dense and continuous GPS networks is now making it possible to monitor the atmosphere. The determination of the amount of water vapour, in addition to providing the correction to the GPS data, will enable scientists to improve the forecast of catastrophic weather events (Naito *et al.*, 1998) and on long time scales to study secular changes in the atmosphere possibly linked to climate change.

GPS has proven to be quite a powerful tool in the study of regional or global crustal motions and deformation. Over the years, the observational approach has progressed going from campaign type, where many stations in a network were occupied by a relatively small number of receivers for short periods of time (from a few days to a few weeks), to continuous networks (CGPS) and approaches in which permanent and temporary occupations are envisaged (Bevis *et al.*, 1997). When discussing the CGPS networks, one has to point out the prominent and unique role played by the IGS, the International GPS Service for Geodynamics which provides the scientific community with high-accuracy orbits, Earth rotation parameters and station coordinates. Since 1992, the IGS precise orbits are routinely available a few days after the satellite observations have been undertaken. This has made a major contribution for the analysis of the GPS data collected by small local or even regional networks.

When applying the GPS technique to the study of sea level variations, one is mainly concerned with high-accuracy determination of vertical movements of the Earth's crust, since tide gauges are attached to the land. Repeated GPS measurements may provide reasonably good information on the vertical motions, but to reach the level of accuracy required for sea level fluctuation studies (vertical rates estimated to an accuracy of better than 1 mm/yr over periods of 5 to 10 years) CGPS is needed. However, in this type of application CGPS is not the only requirement. Before installing the CGPS station next to a tide gauge, several issues must be resolved. The main ones are dealing with the best setting for the GPS antenna in terms of sky visibility and absence of radio disturbances, the stability of the antenna location which includes the study of the geology of the site, selection of GPS receiver of appropriate quality and the vertical connection between the tide gauge and the GPS antenna. In order to ensure that appropriate standards be defined, international working groups in Europe and the Unites States are presently working at identifying the problems and the best way to overcome them.

2.2.3.2 Gravimetry

Iginio Marson

The task of gravimetry is the measurement of gravity, which is the magnitude of the acceleration due to the force of gravity, and of the gravity gradient at the surface of the earth (or near to it). Any mass at the surface of the Earth is affected by the gravitational attraction (newtonian attraction) of the earth and other celestial bodies and by the centrifugal acceleration due to the rotation of the Earth. Gravity hence depends upon the distribution of the terrestrial mass and upon the Earth's rotation. Since both mass distribution within the Earth and rotational parameters are also subject to variations in time, the gravity field of the earth has a spatial and temporal dependency.

The centrifugal acceleration or better the centrifugal potential of a given point can easily be computed from an analytical function, knowing the angular velocity of the Earth. The gravitational potential however must be derived from observations taken at the surface of the Earth or in the space. Therefore gravimetry has a relevant role in geophysics and geodesy. The gravity field can be geometrically described by surfaces of constant potential (equipotential surfaces). The equipotential surfaces are equilibrium surfaces. Since gravity varies with the space because of the internal distribution of masses, the equipotential surfaces are not parallel; an increase in gravity causes a convergence of the equipotential surfaces. One particular equipotential surface is of interest for sea level studies: that is the equipotential surface denoted as the geoid. This is the equipotential surface which approximates to the mean sea level surface at a given epoch. The geoid serves as a reference surface for defining the height system. The morphology of the geoid is space dependent (since it depends upon the internal distribution of masses) and also time dependent, since the distribution of masses in the interior and at the surface of the Earth changes with the time. Both spatial and temporal variations of the geoid are of peculiar interest in studies of sea level changes. Even if the geoid does not coincide with the sea level surface, it is the dominant contributor to the sea level morphology. Over the oceans the geoid can also be efficiently computed from satellite altimetry data, while over the land the measurement of the gravity field is still the prominent method.

Time dependent gravity variations are important in the study and comprehension of phenomena leading to crustal deformation. The study of crustal deformations plays a key role in the determination of mean sea level changes. A crustal deformation process implies a variation of the position (co-ordinates and height) and a variation of the gravity field. This last because the gravity field is directly affected by the variation of the position of the measuring point (mainly of the vertical component) and because crustal deformation is associated with changes in the density field in the Earth interior (due to viscoelastic deformation, pre-seismic dilatancy, dislocation or transfer of internal masses). Therefore, the combination of gravity and position changes allows the computation of changes of the potential and can provide important information on the dynamics of the phenomena. Moreover, gravity changes can be used to validate observed height variations. It is therefore important to measure simultaneously the position and the value of the gravity acceleration at a selected site.

Gravity measurements can be undertaken by means of two types of instruments: absolute gravimeters and spring or superconducting relative gravimeters. Modern absolute gravimeters are based on the ballistic method. In this measuring scheme, an object, typically a cube corner, is either just dropped (free fall) or launched vertically (symmetrical free fall) in a vacuum chamber. A Michelson interferometer, based on the radiation beam of an He-Ne laser stabilised on an iodine absorption cell, measures the position of the cube corner during the flight, while a time interval counter and an atomic frequency clock (rubidium) are used to measure the related time intervals. Since the measurements are made with respect to meteorological standards for length and time, the measurements are absolute. Therefore in contrast to other techniques, absolute measurements are a highly suitable tool for long term studies. According to several experiments performed in the last years, the precision achievable nowadays is 1 μGal (1 μGal = 10 nm/s^2) whereas the accuracy reaches the level of 3 μGal (Marson *et al.*, 1995).

During an absolute gravity campaign, some selected sites on pillars or stable ground are occupied by one or more absolute gravimeters for one or two days. In this time frame, one hundred (symmetrical rise and fall instruments) or two-four thousand (free fall instruments) measurements are taken. The final result, corrected for time-dependent gravity effects (such as earth tide, ocean loading, air pressure, polar motion etc.) is referred to the effective height of the particular gravity meter. To standardise the result, the measured g value should be reduced to the height of 1 m, in order to minimise errors due to the vertical gradient of g (which has to be measured at each site).

In a spring-type relative gravimeter the gravity force acting on the test mass is balanced by the restoring elastic force of a spring which holds the mass itself. In the modern high sensitivity instruments, the equilibrium condition is realised by the equality of the moments of the gravity force and of the spring force with respect of the pivot point of the arm which holds the test mass. Spring gravimeters are used to determine the vertical gradients at absolute stations by differential measurements as well as to tie the tide gauge site to the absolute stations, and to realise microgravity network for local studies. For small gravity differences accuracies of 1 to 3 μGal are achievable (Becker *et al.*, 1995). Spring-type gravity meters also can be used to measure the gravimetric Earth tide signal, even if their drift characteristics prevent their use for long-term projects.

The measuring scheme of the superconducting relative gravimeters is based on the magnetic levitation of a sphere due to the magnetic field generated by a current flowing in a superconducting coil which balances the gravity field. The amount of current necessary to keep the sphere in the reference position is related to variations in gravity field. Owing to the properties of the superconducting coil the instrument should be practically drift free and can be operated continuously for long periods. Because the instrument is a relative one it requires a proper calibration which can be performed in-situ (Richter *et al.* 1995). The performances of these instruments allow the detection of the time-dependent gravity characteristics at the measurement site.

In a long-term project one aims at the detection of a signal of low amplitude (of the order of few μGal per year) and long period (of the order of year or years) affected by a noise which can have an amplitude larger than the signal sought and periods ranging from a few seconds (miscroseismic noise) to several months (water table). An appropriate measurement strategy can be helpful in attempting to increase the signal-to-noise ratio. To this end, it is worth recalling that the sampling rate of the absolute gravity meters ranges from 2-3 minutes in case of symmetrical rise and fall to

10-20 seconds in the case of the free-fall ones. If data are acquired continuously at the highest sampling rate for, let us say, two days, the components of the shortest periods can be detected directly from the absolute measurements. So one could safely state that the noise components with periods ranging from a few minutes to half a day can be detected and removed from the data just by analysing the absolute data set. If the observation time is increased to weeks or months, it is possible, at least from a theoretical point of view, to also detect components with longer periods. The typical measurement strategy is however of the first kind (sampling rate of 10-20 seconds and observation period of one-two days) which limits the frequency range of the noise components which can be detected. In any case, it is obvious that the most critical components are those of longer periods. The most efficient method of dealing with these noise components requires the use of a superconducting gravity meter.

The sampling rate of a superconducting gravity meter is higher than that of the fastest absolute gravimeter so that it has a clear advantage in the high frequency range. A superconducting gravity meter is a relative instrument, which requires a calibration, and might be affected by a certain degree of drift. For these reasons, a superconducting instrument is an excellent instrument to measure gravity changes with periods from seconds to year(s), but not for long-term studies, at least for the very long-term ones. Both calibration and drift characteristics could be assessed by frequent co-locations of absolute and superconducting gravity meters. There is a mutual interest in the co-location of these two type of instruments: the absolute one can calibrate and assess the drift rate of the superconducting meter, while the latter can measure the noise components which result from the inadequate sampling interval of the absolute gravity meter.

2.2.4 Satellite Altimetry

Anny Cazenave

Satellite altimetry provides absolute sea level variation measurements with high and homogeneous spatio-temporal coverage. Two different techniques are used to arrive at observations of sea level: radar altimetry and orbit determination. The radar altimeter measures the height of the satellite above the instantaneous ocean surface and orbit determination provides the distance of the satellite relative to the centre of mass of the Earth, usually expressed as the height of the satellite above a reference ellipsoid (Figure 2.1). The difference between these two measurements is the sea level relative to the reference ellipsoid. This quantity is the sum of two contributions: the height of the geoid (i.e., the equipotential surface of the Earth gravity field) above the reference ellipsoid and the height of the instantaneous ocean surface above the geoid. It is the latter contribution which represents the oceanographic signal and is of interest for sea level investigations. The time-variable sea level height is typically measured at 1-s intervals along the satellite tracks. Altimeter satellites are generally placed on repeat orbit so that after a few days satellite ground tracks exactly superimpose and a global coverage of the Earth is realised during an orbital cycle. Thus the 1-s sea level height measurements may be averaged in space and time during an orbital cycle which allows the global mean sea level to be monitored at regular time steps, i.e., the orbit repeat period.

Half a dozen altimeter satellites have been launched during the past 20 years. However only the most recent altimetry missions (ERS-1/ERS-2 and Topex-Poseidon, launched respectively in 1991, 1995 and 1992) provide sea level height measurements with sufficient precision to allow detailed and precise studies of the variable ocean circulation and open new perspectives in measuring global sea level changes of climatic origin. For previous altimetry missions, errors in the satellite altitude above the reference ellipsoid (called the radial orbit error) were the major limitation for precisely measuring mean sea level variations. This problem has been overcome with Topex-Poseidon, the first altimetry mission specifically designed and conducted for studying the large scale ocean circulation (Fu *et al.*, 1994, 1996; see also the collection of papers in two special JGR issues, vol. 99, Dec. 1994 and vol. 100, Dec. 1995). An unprecedented orbital accuracy is achieved for Topex-Poseidon, owing to a number of improvements in the mission design (multiple on-board tracking systems, in particular the DORIS system) as well as to considerable efforts in providing updated versions of the force model used to compute the satellite orbit.

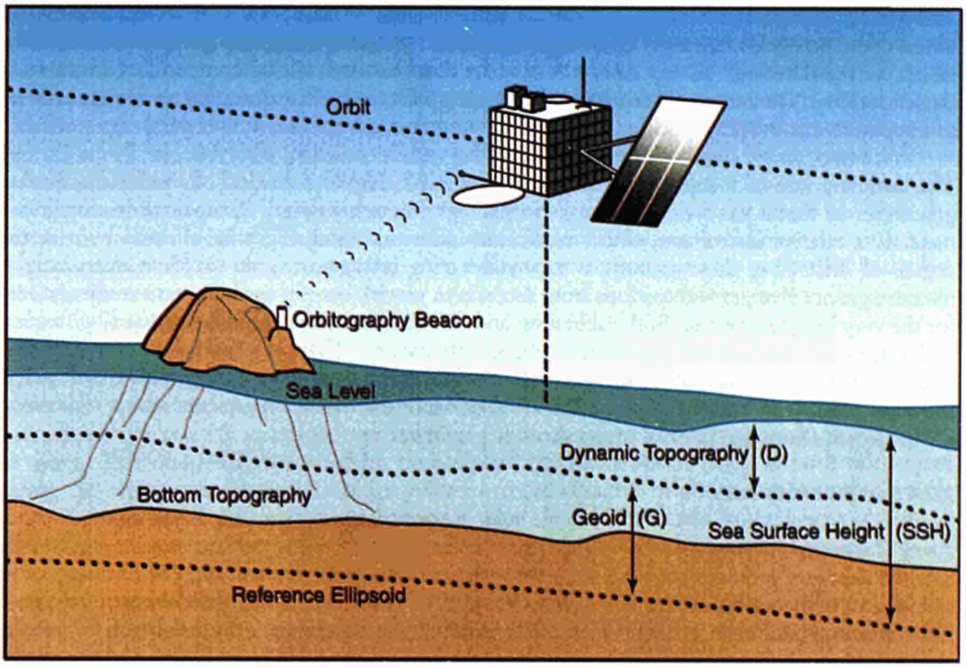

FIGURE 2.1. Principle of altimetric measurement.

The current precision of a single sea level height measurement with Topex-Poseidon is ~2 cm, a significant achievement compared to previous missions. When averaged ocean-wide, the sea level height precision is at the mm level. To arrive at this precision level, much care must also be taken to correct the altimeter measurements for instrumental and signal propagation errors (i.e., ionospheric, dry and wet tropospheric delays) and sea state bias. Moreover the sea surface height measurement needs to be corrected for ocean tide and pole tide to show up ocean dynamics and sea level change signals.

Satellite altimetry provides 'absolute' sea level measurements since these are referred to the Earth's centre of mass (or equivalently to a reference ellipsoid) and thus are independent of vertical land movements. However, any vertical displacement of the sea bottom must be corrected for. This is done indeed for the solid Earth tide and the elastic response of the solid Earth due to ocean tide loading. Another contribution of lower amplitude arises from post-glacial rebound, an effect that is not in general taken into account in satellite altimetry studies.

2.2.5 Airborne Laser Altimetry

A. M. Cocard and Hans-Gert Kahle

Airborne laser altimetry is mainly aimed at the determination of the sea level in coastal areas to bridge satellite altimetry of the deep sea with coastal tide gauge stations. It is an independent method to monitor the sea surface height at the dm to cm level. Airborne laser altimetry allows a validation of the spaceborne radar altimetry results (from ERS1/ERS2).

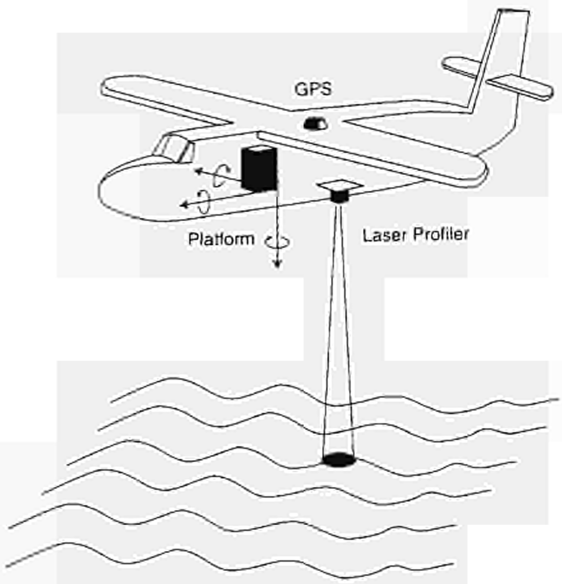

FIGURE 2.2. The laser profiler system including three sensors: GPS, Laser, and INS Platform

The laser profiler system installed on board the aircraft comprises a GPS receiver, a gyro platform and the laser profiler (see Figure 2.2). Basically airborne laser altimetry is similar to the spaceborne concept. In both techniques a good knowledge of the sensor position is required. The problem of highly accurate orbit determination of ERS1 or ERS2 corresponds to the problem of the DGPS-based recovery of the aircraft trajectory which can only be achieved by including differential carrier phase measurements on both frequencies L1 and L2 in the processing of the co-ordinates. Therefore, additional ground-based reference stations are necessary. The laser profiler measures the distance from the sensor to the sea surface at a high frequency (typically 100 - 2000 Hz). Since the laser is strapped to the aircraft an additional inertial sensor records the attitude angles. Thus the 3-dimensional vector from the laser sensor to the sea surface can be recovered at a high precision. By adding this vector to the position vector of the laser sensor deduced from DGPS the co-ordinates of the reflector point at the sea surface are obtained. In both systems, airborne and spaceborne, the data have to be corrected for tidal effects.

The first test flights with an airborne laser system were carried out in 1995 over the Ionian Sea (Greece). One flight path coincided with an ERS2 track. This allowed a direct comparison of spaceborne and airborne heights along a profile. Figure 3 shows the good agreement between both methods. If the airborne survey is designed to allow for crossing points with satellite tracks additional comparisons are possible.

An assessment of the accuracy obtained can be made by different means. As shown in Figure 2.4, a repetition of profiles is a good but expensive option. A more efficient way to obtain control is to use cross-over points. During the planning, special care was taken to obtain a good distribution of such points within one mission and between paths flown on different days. The results obtained in the above mentioned campaign showed that for 72% of all the crossing points the height difference is below 10 cm and for 96% below 20 cm (68 crossing points in total).

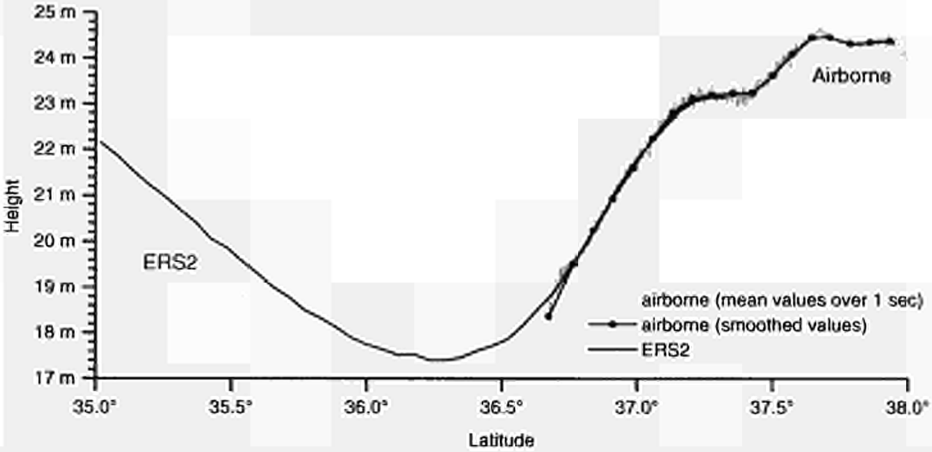

FIGURE 2.3. Comparison of the ellipsoidal height of the sea surface along a profile over the Ionian Sea (Greece) obtained by spaceborne radar (ERS2) and airborne laser altimetry.

Operating an airborne laser system is normally only possible during dedicated missions over small areas compared to the world-wide coverage of satellite altimeter data. Therefore, the availability of airborne systems is restricted and cannot compete with satellite altimetry. Nevertheless dedicated missions allow an independent comparison and are able to collect data over the near coastal areas where satellite altimetry fails and thus allow linkage of the tide gauge measurements with satellite altimetry results.

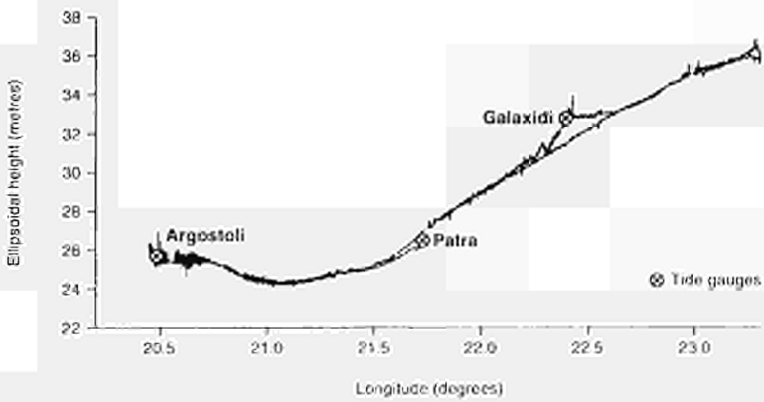

FIGURE 2.4a. Reconstructed ellipsoidal heights of a particular profile across the Gulf of Patras and the Gulf of Corinth. The flight started and ended at the airport of Argostoli. The ellipsoidal heights of the tide gauges which were overflown are shown by ⊗.

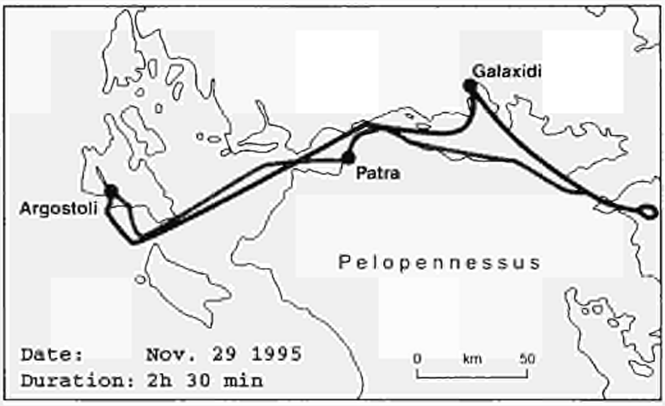

FIGURE 2.4b. The location of the flight track.

2.3 SEA LEVEL DATA ASSESSMENT

Hans-Peter Plag

In the past few decades, several databases with sea level and sea level related observations have been compiled. In this section, we first discuss the relative sea level data derived from various palaeo water level indicators. We then turn to the observational period, which began roughly 200 years ago. Here we constrain the discussion to the monthly mean sea level data derived from tide gauge observations. Finally we comment on the climatological and oceanographic databases relevant to the study of sea level variability.

In the assessment of the databases, the geographical distribution and the temporal span of the records are critical parameters for the global representativeness of a given database. Equally important are data inhomogeneities, which constitute major methodical obstacles when analysing a historical data set. These inhomogeneities are non-geophysical changes which reflect changing instrumental accuracies, environmental conditions, observational practices and analysis routines. A data set is considered "homogeneous" if it is free of such artificial contaminations, which may be creeping or more sudden (Karl *et al.*, 1993; Jones, 1995). Obviously, inhomogeneities are particularly harmful when long term trends are studied.

To promote to the largest possible extent studies of sea level variability and its relation to climate variability, it is required to guarantee free access to relevant data archives for all interested scientists. Relevant archives include meteorological and climatological data bases, oceanographic data, crustal movement observations particularly those acquired with space-geodetic techniques, monthly mean sea level or hourly values derived from tide gauge observations, and satellite altimetry observations or other remote sensing data of the ocean.

2.3.1. Relative sea level data of the last 20 kyrs: the pre-observational period

For the pre-observational time period, former positions of relative sea levels are deduced with the help of morphological, geological, biological or archaeological indicators (see e.g. van de Plassche, 1986; Devoy, 1987; or Pirazzoli, 1996 for overviews). Several research groups have compiled regional or global databases of relative sea level changes over the last 20 kyrs, comprising various types of samples used to derive the timing and location of former sea levels (e.g. Tooley, 1976, 1978, 1982; Newman, 1986; Pirazzoli, 1991; Plag, 1995, Svensson, 1996; Shennan, 1989; Shennan and Woodworth, 1992). Some of these efforts were coordinated in large international projects such as the IGCP Projects 61 and 200. Despite all the efforts, there is still a number of open questions related to the observational evidence used to construct former RSL as well as the concepts inherent in the interpretation of the data (Plag *et al.*, 1996). In particular, the uncertainty in the relation between a given indicator and the local to regional palaeoenvironment in which it is formed constitutes a serious problem in sea level studies. Plag *et al.* (1998a) point out that the geographical distribution of the globally available RSL samples is strongly biased towards the margins of the Laurentide and Fennoscandian ice sheets, i.e. towards areas raised with respect to the former sea levels. They also noted that the overwhelming majority of the samples is given for sea level heights of ± 5 m (with respect to present-day sea levels), and for times less than 12 kyrs BP. More than half the samples are dated for times less than 6 kyrs BP. The RSL samples collected in the larger databases are compiled from a large variety of sources. Consequently, they appear to be rather inhomogeneous. Uncertainties in sea level height and time, if given at all, are often derived in an approximated and subjective way. The reference level for a specific indicator (e.g. highest high water, mean sea level, lowest low water) is not always clear and samples referring to different reference levels may be compiled in one local curve. Moreover, many of the samples are dated with radiocarbon method and corrections for the radiocarbon time scale may be applied inhomogeneously. On the one hand, this clearly indicates the need for a consolidation of the RSL database into one homogeneous, quality-controlled database. Such a database easily accessible for researchers would allow the investigations of past sea level changes in a more coherent way and to determine the gaps and deficiencies of the databases. On the other hand, the inhomogeneous sample distribution in time and space calls for dedicated field work to close the most obvious gaps.

2.3.2 Monthly mean sea level data: the observational period

Within the observational period of the last approximately 200 years, sea level observations have been dominantly carried out by coastal tide gauges, which may have been operated in a purely manually mode or fully automated. Within the last 7 decades, the Permanent Service for Mean Sea Levels (PSMSL) has built up a global data base of monthly mean sea levels derived from these coastal tide gauge observations and thus compiled one of the most valuable climatological data bases. Currently, the PSMSL database contains more than 1700 records with up to almost 200 years of data for the longest records. However, most of the data originate from the time period between 1950 and today and the geographical distribution is strongly biased towards a few coasts on the Northern Hemisphere. Most of the long records originated from ports in Europe and the North American coasts.

The PSMSL classifies the mean sea level data into either 'Metric' or 'Revised Local Reference (RLR)' data. The Metric data set contains all submitted data as they are. The records or part of records classified as RLR are referred to a local reference with the help of a detailed continuous history of the relationship between the tide gauge and the tide gauge benchmark (TGBM). A surprisingly high proportion of mean sea level data, even in Europe, is not in the RLR data-bank (Woodworth *et al.*, 1990). This is usually because either the essential bench-mark information has not been provided by the national authorities or obvious errors have been made in compiling the bench-mark information e.g. after port reconstruction. In some countries, tide gauges are referred to a national levelling datum. However, this practice may introduce spurious jumps of mean sea levels caused by re-levelling or readjustments of the national levelling networks. Therefore, records from such countries (e.g. The Netherlands) are not included in the RLR data set, though they may be of high quality.

A simple check to assess the quality of the monthly mean sea level data is to compare two stations from the same coastal area with each other. In Figure 2.5, the PSMSL records for stations from the Norwegian coast North of 66°N are compared to the record from Tromsø. It is interesting to note that the tide gauges at these stations were replaced with new technology (digital tide gauges) in 1989-90. For each station, the difference to Tromsø clearly displays the inhomogeneity in the records due to this shift of technology, i.e. after 1989, all records show a significantly lower variability. Thus, the high intraseasonal variability of ± 200 mm in the difference between two stations prior to the technological improvements is most likely due to problems of the tide gauges or in the data handling. For some records (e.g. HAM, NAR), small offsets are discernible,

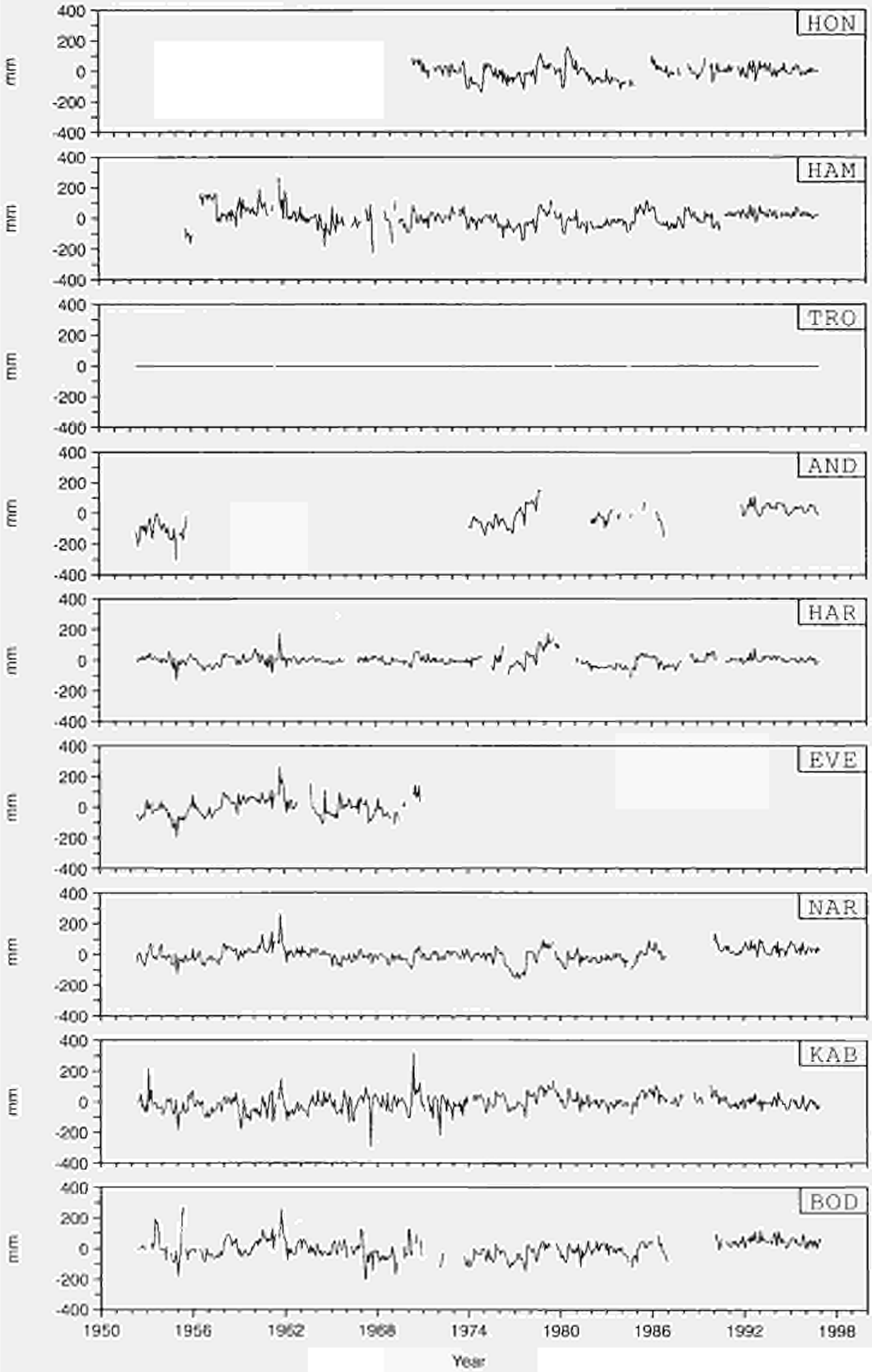

FIGURE 2.5 (caption overleaf)

Intercomparision of monthly mean sea level records. For the full names of the sites, see Table 1. For each station, the monthly differences to Tromsø (TRO) are displayed. To avoid any affect of the main periodic constituents (annual and semi-annual cycle), the monthly differences were computed from residual time series after the annual and semi-annual constituents and a linear trend were removed from the individual records.

particularly those due to the shift of equipment. Spikes found in only one record (like the one at 1970 in KAB) are indicative of data errors in that record. Similar studies for the Mediterranean, Baltic Sea and North American tide gauges reveal the same result with an inter-station variability larger than the expected one for most older, analogue parts of the records. In a careful reprocessing, the Norwegian Mapping Authority, which is responsible for the Norwegian tide gauges have managed to identify and remove many of the data errors and jumps from the records displayed in Figure 2.5. Thus, the effort of reprocessing the data can greatly improve the data quality. Such a reprocessing can only be done by the national authorities, who have access to the original observations and ancillary information. The present quality of the monthly mean sea level data certainly can and should be improved.

In order to improve the distribution and quality of sea level information around European coasts, Baker *et al.* (1997) have proposed a network of key European tide gauges for which there would be a unified system of quality control and data banking of high frequency sea level data, together with the necessary complementary geodetic information from GPS and absolute gravity. This would provide a strategic European sea level network and would be a regional densification of the Intergovernmental Oceanographic Commission's Global Sea Level Observing System (GLOSS). Within the COST Action 40 (EOSS), emphasis is also on defining the boundary conditions necessary for the improvement of both, the quality of the presently operational tide gauges and the sea level data base.

2.3.3 Oceanographic and climatological data sets

Oceanographic data sets have been compiled for climatology as well as temporal variability of, for example, sea temperature and salinity. Reanalyses of meteorological observations of the past several decades have been carried out and have resulted in data sets providing information on the atmospheric forcing of sea level variations. However, the available data sets are of varying quality and all have their specific advantages and disadvantages. Thorough analysis of the existing data sets is urgently needed in order to assess the data quality based on a prescribed, as far as possible homogeneous, and structured method. Assessing the data quality as well as the applicability and limitations of the various data sets for global change studies would help to ensure an optimal use of the available data and to avoid misinterpretations or waste of the available resources. Moreover, the identification of observational gaps both in space and time would help to direct the scarce research resources to those areas with the highest potential for furthering our knowledge and understanding.

2.4 ANALYSIS AND INTERPRETATION OF THE DATA

2.4.1 Past relative sea level changes

C. Romagnoli

On the geological time-scale, records of past sea levels showed that there are globally coherent connections between sea level and climatic/astronomic conditions (Chappel and Shackleton, 1986; Emery and Aubrey, 1991; Edwards *et al.*, 1993). If major climatic changes (such as the ones which have occurred since the beginning of the last deglaciation, e.g. about 18 kyrs ago) imply worldwide RSL changes, it still debated whether, within the same time interval, minor sea level fluctuations may also have occurred, due to minor global or localised climatic oscillations (Pirazzoli, 1996). Several independent and complementary geologic indicators found on continental margins supported, for instance, a stepwise postglacial sea level rise, which seems connected with submillennial-scale deglaciation events (Anderson and Thomas, 1991; Locker *et al.*, 1996). Recognition of short-term RSL fluctuations has major palaeoclimatic and palaeoenvironmental implications and should help the refinement of models addressing linkages within the ocean, ice and climate systems.

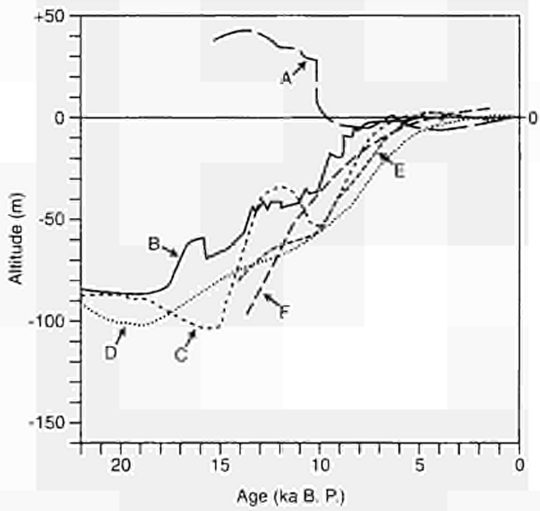

FIGURE 2.6. summarises some available information on local sea level changes for selected sites in Western and Southern Europe, derived from radiometrically dated markers, which enabled the reconstruction of the RSL curves since the last glacial maximum (18-20 kyrs BP; see Pirazzoli, 1996 for details). Apart for curve A, which is derived from an ice-margin area and reflects glacio-isostatic rebound, curves B to F, from outside ice-margin sites, show for the last deglaciation a fast and continuous RSL rise, including periods of more rapid pulses, which are related to accelerated melting of ice caps (Bard et al., 1996; Locker et al., 1996). The average rate of RSL rise between the lowest level reached by the sea at the time of the glacial maximum and the level of the so-called Climatic Optimum of about 6,000 yrs B.P. (which lies close to the present-day sea level) is comprised between 6 and 12 mm/yr, which is substantially faster than either the rate for the last 6,000 yrs rise or that for the past century; during the past 6,000 yrs the rate of sea level rise has decreased sharply, (although the onset of this decrease was not globally synchronous; Gornitz, 1993), with an average rate of 0.6 mm/yr.

Diversities among the curves of RSL for the last deglacial period may be partly explained by different isostatic and tectonic conditions, but also by differences in methodological approach (for example, the approach to calibration of radiometric dating), by the use of more or less reliable sea level indicators, by inadequate or discontinuous vertical resolution or even by the absence of estimation of uncertainty margins (Pirazzoli, 1996; see also section 2.3.1). The statistical fitting of RSL rise curves to a scatter of dated samples, without taking into account adequate uncertainty ranges, for instance, tends to produce a smoothly rising curve and could actually obliterate evidence of minor oscillations during the marine transgression of the last deglacial (Bird, 1996).

On the long-term trends, the rate of relative change of both sea and land movements is similar within an order of magnitude. Generally, sea level fluctuations show relatively high rates combined with a typical oscillatory character, while crustal motions may display (over defined geological intervals) low-rate movements which are generally in a constant direction. In the Late Quaternary, sea level fluctuations generally exceeded the other components (at least on mid-latitude margins, far from areas affected by post-glacial rebound) and occurred in the same depth range (from a few metres above the present-day sea level to about 120-130 m below). For this reason, crustal vertical movements, such as the ones due to regional tectonics or thermal/tectonic subsidence, in most cases did not modify the trend of coastal displacement within each glacio-eustatic cycle, but altered the long-term trend of the RSL at a multi-cycle time scale, as reflected by the stratigraphy of continental margins or by the vertical distribution of sea level markers (Chiocci *et al.*, 1997).

The range of rates for crustal movements may be represented, on the other side, by short-term, non-linear or alternating vertical movements such as the ones due to volcano-tectonic deformations near active volcanoes or to episodic, high-magnitude seismic events. The latter may enhance the rates of crustal movement or produce sudden coseismic vertical displacements of several decimetres opposite to the long-term tendencies, as recognised in wide sectors of the Central-Eastern Mediterranean (Pirazzoli, 1986; Stiros *et al.*, 1994; Pirazzoli *et al.*, 1994) and recorded through large fluctuations of the RSL. To assess the long-term trend of uplifting or subsidence at the local/regional scale a closely-spaced grid of reliable markers for the past RSL is needed; moreover, average rates of crustal vertical movements, and associated RSL changes, should be referred only to the time period relevant to the field observations.

2.4.2 Tide gauges

Hans-Peter Plag

During the last decades, the global data set of monthly mean sea levels provided by the PSMSL has been the basis of a larger number of sea level studies aiming at a detection of a global or regional sea level rise or even an acceleration of the trend in sea level. The results and main conclusions of the majority of these studies with respect to sea level rise are summarised in the recent assessment of the IPCC (see Warrick *et al.*, 1996). Therefore, here we will concentrate on the methodology used to determine a global sea level change from tide gauge records and on inherent problems in the approach.

When using tide gauge observations to determine the change in the global ocean volume, basically three steps are required: (1) the determination of local trends at each tide gauge, (2) the correction of the effects of crustal vertical motion, (3) the global averaging of the local trends. In what follows, we will discuss each step separately. The result of the three steps, if they can be achieved at all, will be a change in global volume, which then needs to be separated into steric (pure volume changes) and non-steric (mass) changes of the ocean. This difficult and still unsolved problem and the difficulties in explaining these changes (i.e. the ocean's mass and heat balance) are discussed in other sections.

The accuracy of the trends in relative sea level determined from tide gauges not only depends on the quality of the tide gauge recording and the maintenance of the relationship between tide gauge and TGBM but also on the length of the series of observations. The interannual to interdecadal variability of coastal sea level is of the order of several cm and more. In many regions, the long-term linear trends over the same time interval are smaller or of the same order as this variability. Therefore, trends determined from shorter records may be strongly biased by the interannual to interdecadal variability. In fact, trends determined from records with spans from years to a few decades show a wide scatter due to this effect of the decadal variability. This has been pointed out by a number of authors (e.g. Pirazzoli, 1986; Gröger and Plag, 1993; Zerbini *et al.*, 1996). Baker et al (1997) showed that an accuracy in local trends of (0.5 mm/yr requires a record length of at least 30 years. Climatologically induced trends in sea level are typically of the order of 1 mm/yr, and to determine these trends reliably requires much longer records.

Already Sturges (1997) noted the large-scale coherency of the inter-annual to multidecadal sea level variability. Based on the same notion, Sjöberg (1987) suggested a method for correcting the effect of the interannual to interdecadal variability on trends determined from shorter records. Using a high-quality record with a long span (> 50 years) as base station for a given region, the interannual to interdecadal variability can be eliminated from shorter records by considering the monthly differences between a given record and the base station. In this way, the bias due to the interannual to interdecadal variability can be removed from trends determined from short records. It should be pointed out that the success of this method depends on the appropriate selection of the base station.

Besides the interannual to interdecadal sea level variability, local trends are strongly affected by tectonic or other vertical crustal movements and trends due to post-glacial rebound. The latter includes both, a change in geocentric sea level and vertical crustal movements. A large fraction of the ocean's coast is in deformation areas associated with plate tectonics (see e.g. Figure 6 in Stein, 1993). In most of these regions, the tectonically caused vertical crustal movements are dominated by short spatial scales of the order of 30 to 100 km (see e.g. Emery and Aubrey, 1991). In most analysis of the tide gauge data, tectonic movement has not been corrected for. The usual treatment is to exclude stations from knowingly unstable coastal areas (see below).

In some studies aiming at the detection of the recent trend, the geological trend determined at nearby sites has been used to correct for both, long-term tectonic and postglacial trends (e.g. Gornitz and Lebedeff, 1987). In most recent studies, the effect of post-glacial rebound is corrected for by using geophysical models (e.g. Peltier and Tushingham, 1989, 1991; Trupin and Wahr, 1990; Mitrovica and Davis, 1995; Douglas, 1997; Peltier and Xianhua Jiang, 1997). There are, however, large uncertainties in the geophysical model prediction of the present-day post-glacial sea level trends, which are mostly due to uncertainties in the vertical viscosity structure of the Earth's mantle (see e.g. Plag *et al.*, 1998b, for a discussion). Mitrovica and Davis (1995) concluded that the uncertainties of the geophysical models results in an uncertainty of the corrected global sea level trend of 0.5 mm/yr. However, particularly in regions closer to the former ice sheets, the uncertainty in the predicted present-day post-glacial trends in sea level may be much larger than that, as is demonstrated in Table 1. Using two models with only a small the difference in the mantle viscosity results in differences in the post-glacial signal of up the 1.8 mm/yr at Arctic tide gauges. Using a full suite of the available geophysical models would result in much larger discrepancies. It is interesting to note that most of the stations listed in Table 1 are located at tectonically stable areas. Nevertheless, the vertical crustal movement rates deduced from these tide gauges turn out to be much larger (of the order of (4 mm/yr) than, for example, similar rates determined from the tide gauges in the Mediterranean (of the order of ± 1 mm/yr; see Zerbini *et al.*, 1996). Plag (1998a) interprets the fact, that most of the vertical crustal rates determined from the Arctic tide gauges turn out to be positive, as a possible indication of elastic uplift due to a reduction in the ice load over Greenland and Svalbard (the latter for the two tide gauges at Ny-Alesund and Barentsburg)

It is interesting to note that all values for a global mean sea level rise determined in numerous studies during the last few decades (see e.g. Table 4 in Pirazzoli, 1996, for an overview) basically rely on the same global tide gauge data set. Nevertheless, the resulting global values scatter over a range from 1 to 3 mm/yr. The differences in the results mainly are due to different data selection criteria and different methods for global averaging. In early determinations of a global sea level rise, simple averages were taken over all local trends determined at all or a subset of the global tide gauge set (e.g. Barnett, 1984). This approach was based on the assumption that most biasing effects due to vertical land movements would globally average to zero. Later studies are also mostly based on averaging over a large number of local trends after correcting these trends for the present-day post-glacial sea level trend (see above).

Peltier and Tushingham (1989) point out that the global average strongly depends on the choice of minimum record length and the particular time interval considered. Gröger and Plag (1993) also point out the strong effect of data selection on the resulting global average. In agreement with others (e.g. Pirazzoli, 1986; Emery and Aubrey, 1991), they concluded that due to the deficiencies

TABLE 1. Trends at Arctic tide gauges.

Station	Long.	Lat.	Beg.	End	N	t	δt	p_1	p_2	r'_1	r'_2	v_{01}	V_{02}
Bodø	14°23'E	67°17'N	1949	1996	482	-3.049	0.5	-2.5	-3.6	-0.5	0.6	2.3	1.2
Kabelvåg	14°29'E	68°13'N	1880	1996	572	-0.962	0.3	-0.5	-1.6	-0.5	0.6	2.3	1.2
Narvik	17°25'E	68°25'N	1923	1996	685	-3.245	0.2	-2.2	-3.5	-1.0	0.3	2.8	1.5
Evenskjær	16°23'E	68°35'N	1947	1970	267	-4.000	0.7	-1.3	-2.5	-2.7	-1.5	4.5	3.3
Harstad	16°33'E	68°48'N	1952	1996	497	-0.053	0.4	-0.9	-2.1	0.8	2.0	1.0	-0.2
Andenes	16°09'E	69°19'N	1933	1996	362	1.538	0.5	0.2	-0.9	1.3	2.4	0.5	-0.6
Tromso	18°58'E	69°39'N	1952	1996	553	0.034	0.3	-0.8	-2.1	0.8	2.1	1.0	-0.3
Hammerfest	23°40'E	70°40'N	1955	1996	471	-0.137	0.4	-1.2	-2.7	1.1	2.6	0.7	-0.8
Honningsvåg	25°29'E	70°59'N	1970	1996	292	2.705	0.5	-1.3	-2.9	4.0	5.6	-2.2	-3.8
Vadsø	29°45'E	70°04'N	1969	1987	206	-2.318	0.6	-2.1	-3.9	-0.2	1.6	2.0	0.2
Vardo	31°06'E	70°20'N	1947	1996	322	-0.755	0.5	-1.6	-3.4	0.8	2.6	1.0	-0.8
Linakhamari	31°22'E	69°39'N	1931	1939	107	-4.377	1.2	-2.2	-4.0	-2.2	-0.4	4.0	2.2
Murmansk	33°03'E	68°58'N	1952	1996	530	1.581	0.3	-2.4	-4.1	4.0	5.7	-2.2	-3.9
Russkaya Gavan	62°35'E	76°12'N	1953	1991	457	-0.851	0.4	-2.8	-4.5	1.9	3.6	-0.1	-1.8
Tuktoyaktuk	132°58'W	69°25'N	1962	1982	156	7.134	1.1	3.0	2.7	4.1	4.4	-2.3	-2.6
Cambridge Bay	105°04'W	69°07'N	1965	1982	121	-3.991	1.0	-3.5	-5.8	-0.5	1.8	2.3	0.0
Resolute	94°53'W	74°41'N	1957	1977	206	-2.883	0.9	-3.2	-4.3	-0.3	1.4	2.1	0.4
Ny-Ålesund	11°56'W	78°56'N	1976	1996	175	-1.373	0.8	-0.9	-1.3	-0.5	-0.1	2.3	1.9
Barentsberg	14°15'E	78°04'N	1948	1996	560	-2.253	0.3	-2.0	-2.6	-0.3	0.3	2.1	1.5

Monthly mean sea level data are taken from the PSMSL data base. N is the number of monthly values in the record. **t** *is the linear term and* **δt** *the standard error.* $\mathbf{p}_1$ *and* $\mathbf{p}_2$ *are the post-glacial sea level signal computed for two models with an elastic lithosphere of 120 km, and upper mantle viscosity of 1 x 10^{21} Pas and a lower mantle viscosity for model 1 and 2 of 2 x 10^{21} Pas and 4.75 x 10^{21} Pas, respectively. The ice model is ICE-3G (Tushingham and Peltier, 1991).* **r'** *is the decontaminated* trend **r'= t-p** *and* $\mathbf{v}_0$ *the tide-gauge determined vertical crustal movement rate* $\mathbf{v_o}$**= - (t-p-r')** *where* **r'** *has been take as 1.8 mm/yr. All rates are in mm. From Plag (1998a).*

of the global tide gauge data set with respect to geographical and temporal distribution, a global value cannot be determined. Plag (1993) investigates the assumptions inherent in the determination of a global sea level rise from tide gauges and finds that these assumption are not justified. An rather peculiar approach is selected by a sequence of studies by Douglas (e.g. 1991, 1997). Here, only long records are used, and from them a small number are selected on the basis of several selection criteria. Ending up with only 17 records, Douglas (1997) determines the average of these local trends to be 1.9 mm/yr and equals this to the global sea level rise.

Plag (1998b), on the other hand, argues that all these approaches do not take into account the physical relationship between changes in the ice load on land and sea level. This relation is described by the hydrostatic sea level equation first given by Farrell and Clark (1976). Due to the gravito-elastic response of the solid Earth to ice-load changes, the resulting sea level changes show a characteristic finger-print depending on where the ice load is decreasing or increasing. Thus, a decrease in the Greenland ice sheet would be associated with a sea level fall in the Arctic and a sea level rise in areas further away from Greenland, while a decrease of the Antarctic ice sheet would result in a finger-print with sea level fall around the Antarctic and sea level rise in areas further north. Plag (1998b) claims that only a rigorous inversion of the tide gauges based on the sea level equation can give reliable results for the global average. This inversion remains to be done.

2.4.3 GPS

Sussana Zerbini
The rather heavy burden of analysing GPS data collected by CGPS networks, or during episodic investigations involving a large number of receivers, has considerably relaxed during the past few years through the availability of the IGS precise orbits and co-ordinates and velocities of IGS

reference stations, as well as through the development of user-friendly software packages for high-precision geodesy.

These recent developments are most important since they represent the necessary step toward contributing to the realisation of the concept of "sustainable monitoring" presented by Blewitt *et al.* (1998). Blewitt and his co-authors, by recalling the importance of exploiting old geodetic data series for identifying and studying long-term geophysical signals, introduce the concept of "sustainable geodetic monitoring" as: The production of geodetic data which will be as useful and amenable as possible for future generations. This concept is of particular importance in the study of sea level changes where high-accuracy vertical rates at the tide gauges needs to be determined. Also, this concept is well in line with the approach adopted by the European Commission in the development of the Framework 5 Programme, where the monitoring aspect and the European contribution to the global observing systems are one of the relevant objectives within the "Global Change, Climate and Biodiversity" Key Action.

Vertical rates are influenced by a variety of environmental factors, the effect of which, only in a few cases, can already be accounted for in the GPS data analysis by relatively accurate models. This is true, for example, for the surface displacements induced by forces of exogenic origin such as solid Earth tides and ocean loading. However, other exogenic effects including storm surges, seasonal sea level changes, regional and local air pressure variations, seasonal changes in the global air pressure distribution, snow cover and groundwater changes, glacial retreat, etc. could be estimated by using observational data and the existing knowledge about the viscoelastic behaviour of the Earth (Zerbini *et al.*, 1998b). Other effects, such as postglacial rebound, the extent of which still demands for further observations and understanding, can be estimated from presently ongoing large or small-scale observation programs. BIFROST and BAYONET are such examples (Bennett *et al.*, 1996; Nerem *et al.*, 1998). Land subsidence can be relevant both inland and at coastal sites and it may vary significantly because of human influence (ground fluids withdrawal) (Zerbini *et al.*, 1997). Secular trends in plate motions can be accommodated by adopting the most recent ITRF (International Terrestrial Reference System) station velocities, although shorter-scale tectonic motions and deformation connected to earthquake cycle will still be present in the data.

At present, the analysis of the data provided by CGPS networks indicates that daily solutions for the vertical component can be achieved at the 5 mm level of accuracy, as it has been demonstrated by Zerbini *et al.* (1998a) for the Medicina and Port Corsini stations of the SELF II project network (Zerbini, co-ordinator, 1996). Medicina is a fiducial site near Bologna, while Port Corsini is located on the Adriatic coast near the city of Ravenna. The analysis of the time series spanning more than two years now seems to indicate that the present length of the time series might not be adequate to identify the vertical rates at the stations reliably (Figure 2.7 a) and 2.7b)) (both sites are affected by subsidence, though of different origin) and that they might be influenced to some extent by the stratigraphic-structural setting of the sites and also by the local micro climates. A spectral analysis of the CGPS series indicates for the stations the presence of significant terms with different periodicity (Figure 8). Both stations exhibit an annual, semi-annual and 100 day periodicity. Also a quasi-biennial term seems to be present in both data series, though the series are still too short (29 months) to definitely claim the existence of this period. While Medicina exhibits a more energetic power spectrum for the annual and semiannual terms with respect to Port Corsini, the quasi biennial and 100-day components are more relevant for Port Corsini (Zerbini, co-ordinator, 1998; Negusini *et al.*, 1998). At both stations meteorological and environmental parameters are being recorded in the attempt to identify the causes for the observed height fluctuations.

These recent findings bring the attention to the fact that to achieve millimetre accuracy CGPS daily heights and better than 1 mm/yr accuracy vertical rates within 5 to 10 years it will be necessary to collect, together with CGPS data, environmental data series which will help to unravel the observed fluctuations in the vertical component. Also, a detailed knowledge of the local geology of the site will be necessary to properly interpret the data.

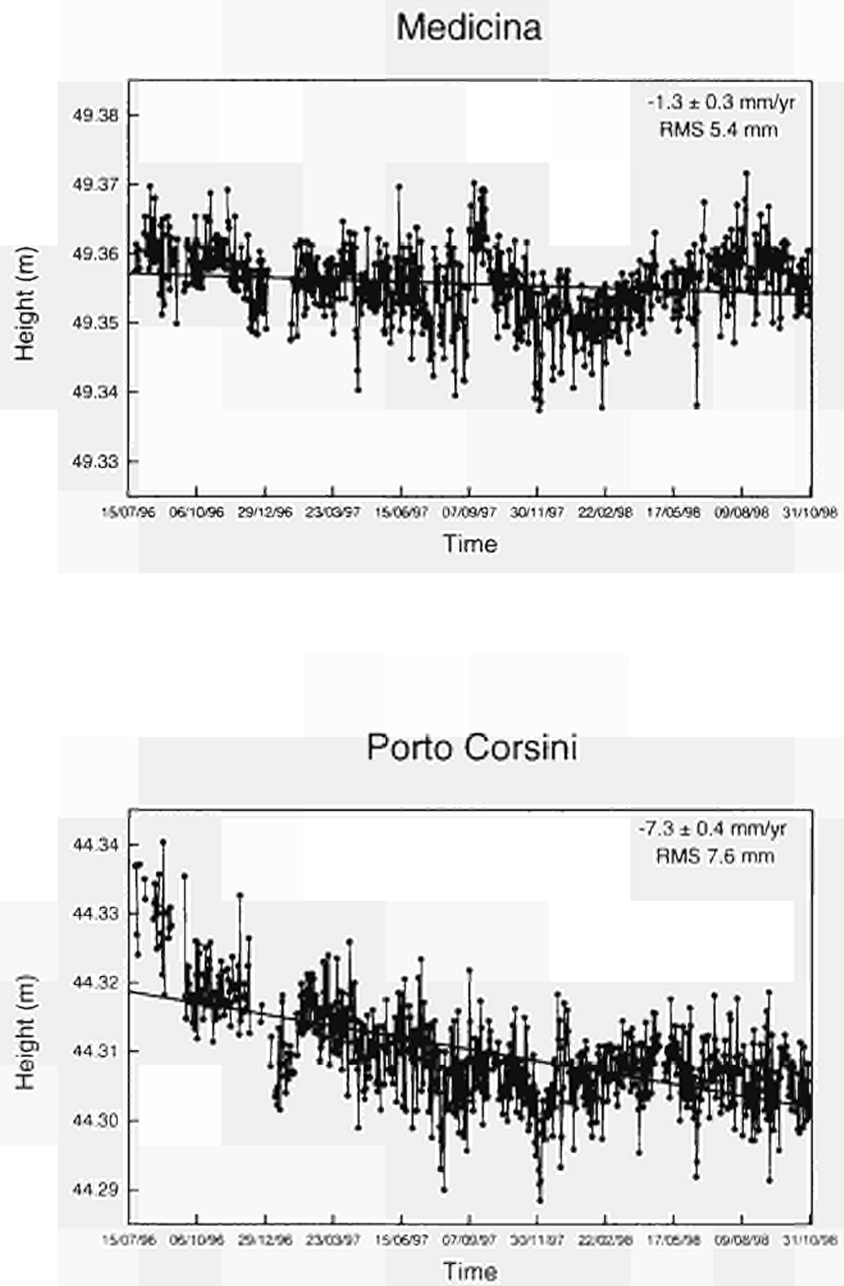

FIGURE 2.7. GPS daily solutions for the Medicina (a) and Port Corsini (b) stations. The linear interpolation through the data provides an estimate for the subsidence rate at the sites.

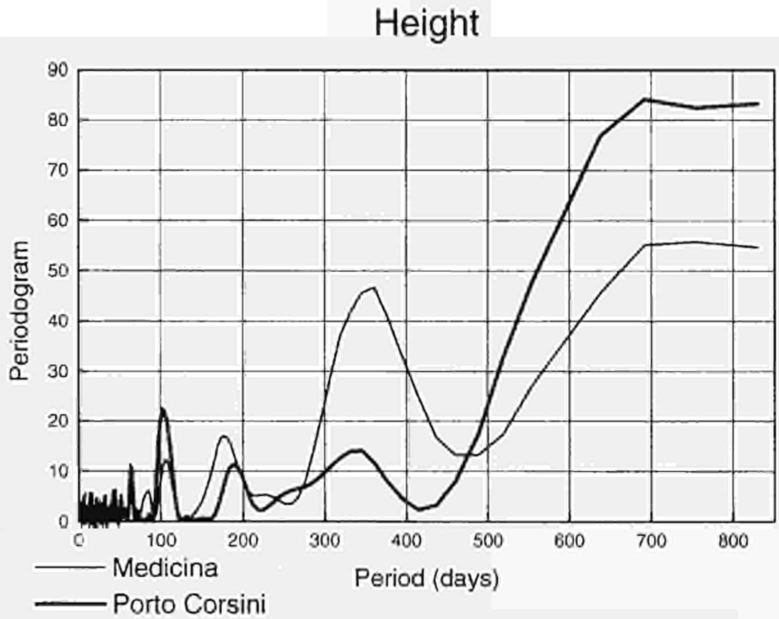

FIGURE 2.8. Periodicity analysis for the Medicina and Port Corsini (thick line) daily height series.

2.4.4 Gravimetry

Iginio Marson

Absolute gravimetry measures (g) by counting wavelengths and measuring time intervals: meteorological standards are used in both space and time measurements. Space geodetic techniques (VLBI, SLR and GPS) use lasers or microwave sources and time intervals to measure the "relative" position of the site. The metrologies of absolute gravimetry and space geodesy are therefore similar. The co-location of absolute gravimetry and space geodesy hence defines a sort of absolute metrology with which it is possible to determine an absolute reference system suitable for long term studies of crustal deformation phenomena. There is no doubt that the meteorological characteristics of the measurements give to absolute gravimetry a leading role in such studies. Always, but particularly when dealing with long term gravity variations, we require high accuracy of the measuring device and high stability of the observation site. As the accuracy requirements increase, the issue of the gravity variability of the observation site becomes more problematic. Several factors can affect this variability in the medium to short periods components, the more important of which are: micro-seismic noise, barometric pressure, water table, solid and fluid tides, loading effects, and polar motion. Some of these could be efficiently modelled or computed if the corresponding physical parameters are given. However in most cases the problem of removing the medium to short period components can be solved only by direct measurements.

The ideal observation site for long term studies, as has been stated above, would be equipped with a permanent GPS receiver, a superconducting gravity meter periodically (two to four times per year) controlled by an absolute gravity meter. The cost for installing and running such a permanent site is relatively high and is affordable only by a few agencies or institutes in the world. Because of their amplitude and period, the gravity effects of water table and air pressure changes are relatively critical. The influence of water table changes on gravity measurements has been carefully studied by M. Delcourt-Honorez (1991, 1995). In these extensive studies, the hydrogeological perturbing effect on local gravity at the Royal Observatory of Belgium in Brussels has been computed by measuring the water level variations at long, short term and due to Earth tidal variations. Several periods have been observed in three aquifers at different depths (-36 m, -62 m and -67 m) with different amplitudes ranging from 0.07 m to 0.18 m. It is interesting to remark that these relatively small water table variations have been detected in the gravity records acquired with a superconducting gravity meter and properly modelled. The total hydrogeological effect is the result of the sum of the gravity effect induced by the water masses in motion and of the gravity effect induced by the land surface displacement. A maximum effect of 4 (Gal has been observed. Gravity effects up to 30 (Gal induced by water-table variations of about 8 m, with seasonal periods, have been also observed (Giorgetti *et al.,* 1987). Long term gravity variation studies need therefore the measurement and modelling of the water table effect. This obviously requires the availability of water wells, knowledge of the physical characteristics of the aquifers and geological structures and efficient mathematical models. If these data are not available, one might make use of the regional water table data, which should be generally available. A superconducting gravity meter, or at least an earth tide gravimeter, could be installed for at least one year at the base station and the acquired gravity data could be used to compute the gravity admittance of the regional water-table data.

The effect of the atmospheric pressure on gravity measurements has been extensively studied by Sun et al (1993, 1995). Atmospheric pressure changes affect gravity through direct gravitational attraction of the air masses and through the induced elastic deformation of the Earth. The authors computed the gravity changes at the surface of a spherical, radially stratified elastic Earth, under the action of the atmospheric pressure by performing a numerical comparison between the mass loading Green functions and the local and regional barometric pressure data. The geographical distribution of the barometric data, extending to more than 1000 km around the station and over 6 layers until about 12 km above the earth's surface is of interest. The conclusions reached by the authors show that the maximum peak to peak amplitude of the direct Newtonian attraction reaches 22.3 μGal at Brussels for an extension of the column load of more than 1000 km. The corresponding admittance is of -0.451 μGal/mbar which is larger than the theoretical value of -0.419 μGal/mbar. Another interesting result is that the atmospheric effects on gravity are affected by the lateral extension of the column load. In fact the admittance of the total gravity effect varies from -0.395 μGal/mbar for loads of 100 km of radius to -0.333 μGal/mbar for loads of more than 1000 km. Regional barometric pressure data over 1000 km around the station are hardly available. Fortunately about 80% of the total effect can be computed using only the first 100 km, which could be an achievable target. Also in this case it could be worthwhile to install a recording gravity meter for at least one year and compute the gravity admittance of the regional barometric pressure data.

2.4.5 Satellite altimetry

Anny Cazenave

The rate of sea level change (for 1993-1996) as a function of geographical location (see chapter 1, section 1.2.4.3, Figure 1.14) shows considerable regional variations ranging from approximately +50 mm/yr to -50 mm/yr. It is interesting to notice that while in some areas sea level is rising (e.g., in the Western Equatorial pacific, North Pacific, South Atlantic), in other regions, sea level is dropping (e.g., Eastern Pacific). The spatial pattern shown on this map is dominated by the regional inter-annual variability of the ocean circulation. Moreover, the positive sea level anomaly seen in the Western Pacific is likely related to the prolonged series of ENSO events which occurred since the early 90s, in particular in 1992 and 1994. On the other hand, the geographical pattern of the SST drift (not shown) shows some striking correlation with that of the sea level, in particular in the Pacific, an indication that regional sea level changes are in part related to heating or cooling of upper ocean. The pattern shown in Figure 1.14 (chapter 1) confirms that sea level variations associated with temperature changes should not at all be uniform, neither in the open ocean not along coastal regions.

2.4.6 Airborne Laser Altimetry

Before retrieving the co-ordinates of the aircraft, all reference stations on the ground have to be tied to the reference frame used for satellite altimetry in order to ensure a bias-free merging with satellite altimetry data.

The kinematic GPS data is processed using precise IGS ephemeris and data from the reference stations. The strategy used to deal with kinematic GPS data, is to process code and phase measurements simultaneously. The basic idea is to perform a least-squares estimation in two steps. In the first step the phase ambiguities are estimated. For each epoch the co-ordinates are eliminated and the reduced normal equation systems are accumulated over the duration of the flight. The advantage of this procedure is that it allows the implementation of the total amount of measurement in the estimation as well as fixing of the ambiguities to integer values. The entire calculation is repeated in the second step, where the integer-valued ambiguities are introduced as known. Here the system is solved for the co-ordinates.

After the recovery of the GPS trajectories the co-ordinates of the sea surface are calculated by adding the vector from the aircraft to the sea surface as measured by the laser and the platform gyros. The ellipsoidal height is extracted by conversion of the geocentric Cartesian co-ordinates to ellipsoidal ones. Different smoothing techniques are applied to eliminate the waves and after applying tidal corrections the mean sea surface is obtained along the paths flown.

Insofar as the problems which remain and may degrade the solution it is worthwhile mentioning the propagation of multipath effects and unmodelled tropospheric errors. The selection of concordant reference systems and reduction parameters for satellite radar and airborne laser altimetry has to be carefully analysed. An important error source is introduced by gyro drifts and offsets in the orientation of the laser beam with respect to the INS navigation frame leading to offsets in the calculation of the Nadir pointing of the laser. Mounting the laser directly on a stabilised (gimballed) platform mitigates these uncertainties.

2.5 TOWARDS PREDICTING REGIONAL AND LOCAL SEA LEVEL VARIATIONS

Hans-Peter Plag

Current models used for the prediction of future changes in sea level in the context of, for example, the IPCC assessments are still based on many simplifications (see Warrik *et al.*, 1996, for example), which currently may be justified by the considerable uncertainties in the quantification of the basic processes affecting the ocean mass balance and volume. However, as these uncertainties can be expected to be reduced in the future, efforts should be made to improve the prediction models particularly by incorporating more feedbacks and by setting up models with better spatial resolution where appropriate.

It must, however, be emphasised strongly that future changes in GOMV, as discussed in the IPCC assessments, cannot be directly equated to changes in relative sea level at a given location. Neither sea level changes due to mass exchange between ocean and cryosphere nor those due to volume changes associated with thermal expansion can be expected to be globally uniform. Mass exchanges between ice and ocean result in deformations of the Earth as well as changes in the geoid both affecting the distribution of ocean water and thus relative sea level. Geophysical models can be used to compute the water distribution and relative sea level if the origin of the mass added to the ocean or the destination of the mass taken from the ocean (basically, i.e. if the ice load history) is known. At present, A/OGCMs can give estimates of the spatial variability in the thermal expansion of the ocean waters due to the combined effect of air temperature, surface pressure, wind stress and ocean currents but these models need to be further developed to include changes in the mass of the ocean. Downscaling from model resolution to the local scale eventually combining the A/OGCMs and geophysical models needs further development using either statistical or nested model techniques. The objective is to provide a sound basis for integrated coastal management particularly in areas with potentially high economic and social consequences. Directly equating future changes in GOMV to regional or local relative sea level changes, as is still often the case even on national level, may direct considerable economic resources into mitigation attempts in areas of relatively low risk, while in other areas a much higher risk may not properly be recognised and timely attempts to mitigate the consequences may not be made.

Moreover, potential hazards for low-lying coastal areas not only arise from possible future rises in sea level, but also will result from changes in the sea level variability due to variations in the atmospheric forcing or the response of the ocean to this forcing. Therefore, studies of sea level variability in relation to changing atmospheric forcing are required on time scales relevant to storm surges, where both an increasing frequency as well as severity may have considerable effects on the anthroposphere. At longer time scales of up to several decades, coupled atmosphere-ocean phenomena dominantly influence climate variability, and studies of atmosphere-ocean interactions at these time scales may provide the urgently needed basis for understanding how future climate changes will affect climate variability on seasonal to centennial time scales.

2.5.1 Wind driven sea level variations in the past and in the future

Eigil Kaas, Hans Von Storch, and Ignacio Lozano

If the oceans were homogeneous and at rest, with a uniform atmospheric pressure field above them and no wind, the sea surface would correspond to the geoid; but actually the mean sea level deviates from the geoid by ±one metre due to forcing as a result of the sun driving the coupled atmosphere/oceanic climate system. A considerable part of sea level variation is directly or indirectly - via currents and baroclinic Rossby ocean waves - related to wind stress, precipitation and atmospheric pressure. The wind/pressure field impacts both the long term mean sea level and the variations around this mean. These variations in sea level take place on time-scales from hours (surges) to decades or even centuries and together with storms and wind driven surface ocean waves (on the time scale of seconds) they are key players in shaping the coastal morphology (see Chapter 3). The present section deals with sea level variations in the North Atlantic/European region on the shortest time scales, i.e. surges as well as with surface ocean waves related to extra-tropical cyclonic activity. In section 3.1 slower variations, i.e. month(s) and longer, in the wind/pressure driven sea level will be discussed.

2.5.1.1 Storms, waves and surges. Extreme events.

Two interesting questions in the context of this volume are:

- how did the storms and wind driven sea level variations change or vary in the past and
- what changes may we expect in future as a consequence of enhanced greenhouse forcing of climate.

In the public debate concerning climate change and increasing emissions of radiatively active trace gases in the atmosphere many are concerned about the possibility of an intensification of extra-tropical storms. Even though the IPCC took a cautious stand in this matter because of lack of evidence, a mixture of indirect evidence (van Hoff, 1993; Hogben, 1994) and controversial scientific statements (Schinke, 1992) created a substantial uneasiness in the public (Berz, 1993; Berz and Conrad, 1994; Greenpeace, 1994). The offshore oil industry in the North Sea was confronted with reports about extreme waves higher than ever observed. The insurance industry organised meetings with scientists because of greatly increased storm-related damage. The northern European newspapers were full of speculations about the enhanced threat of extra-tropical storms in the early part of 1993. For these reasons the EU-supported WASA project was established[1]. The present sub-section is mainly based on results obtained in WASA which focused on the conditions in the North Atlantic/European region. The reader is referred to WASA (1997) for a detailed presentation of WASA results.

2.5.1.2 Past changes in storms and surges

Wind observations in the past 100 years include so many partly unknown inhomogeneities such as changes of measurement scale, change of observer, change of surroundings (e.g. growing or cutting of trees), that they can not be used to study changing storminess (Peterson and Hasse, 1987). Another candidate could be historical weather maps where one can count the number of lows below certain thresholds (e.g. Schinke, 1992; Stein and Hense, 1994). But even the weather maps are inhomogeneous simply because their quality has steadily improved. The improvements are due to the gradually more and better observations and particularly to the introduction of numerical data assimilation systems in the 70s. The more detailed mapping of pressure which is possible nowadays, automatically yields more and deeper lows, a typical example of an inhomogeneity which produce an artificial increase in number of intensive cyclones.

For these reasons one must use other more homogeneous proxies of storminess. An obvious choice is to base these on station air pressure, the time series of which are considered to be relatively homogeneous because more or less the same instrument (mercury barometer) and procedures have been used throughout the entire observation period.

From air-pressure several proxies for storminess may be formed, e.g. the annual (seasonal, monthly) distribution of the geostrophic wind speed derived from three stations in a triangle (Schmidt and von Storch, 1993; Alexandersson *et al.*, 1997). One of the objectives in the WASA project was to collect, digitise, quality control and homogenise station pressure data from a list of approximately 20 stations in the North Atlantic and Northern European Region, and to calculate time series of geostrophic winds from a large number triangles using these data. Figure 10 shows the annual values of standardised 95- and 99-percentiles of geostrophic winds derived from station observations of MSLP using the triangle method. The figure shows that there have indeed been many anomalous events of high geostrophic winds over the British Isles, North Sea and Norwegian Sea in recent years. However, when compared with the values near the end of the previous century they are not extreme or unpreceded. With weaker amplitude the same variation is seen further to the east, over Fennoscandia and the Baltic. Since inhomogeneities in the pressure record from just one of the stations will lead to corresponding inhomogeneities in the geostrophic wind speed one may instead analyse the annual (or seasonal) distribution of the pressure variations, after suppressing the non-synoptic variations by means of a digital filter (Schmith, 1995; Kaas *et al.*, 1996; Schmith *et al.*, 1997). Even though such data are more indirect they are homogeneous proxies. Figure 2.10 shows the temporal variations in the absolute 24 hour pressure tendency at the station of Torshavn at the Faroe Islands in the middle of the Atlantic storm track. Note, the correspondence between the ups and downs in the low pass filtered curves in Figures 2.9 and 2.10. There are, however, certain differences, particularly in the end of last century, where geostrophic wind anomalies were very high, while the 1% and 10% exceedance levels of 24 hour pressure tendencies were more normal. This might be due to inhomogeneities in the geostrophic winds or simply that there is no one to one relationship between pressure tendency and wind speed.

[1] The full title of the WASA project is "Impact of storms on waves and surges: Changing climate in the past 100 years and perspectives for the future". The project was carried out in the period 1994-1996. A description of the project can be found on the Internet (http://w3g.gkss.de/G/Mitarbeiter/storch/wasa.html). WASA was co-ordinated by Dr. Hans von Storch (MPI/GKSS)

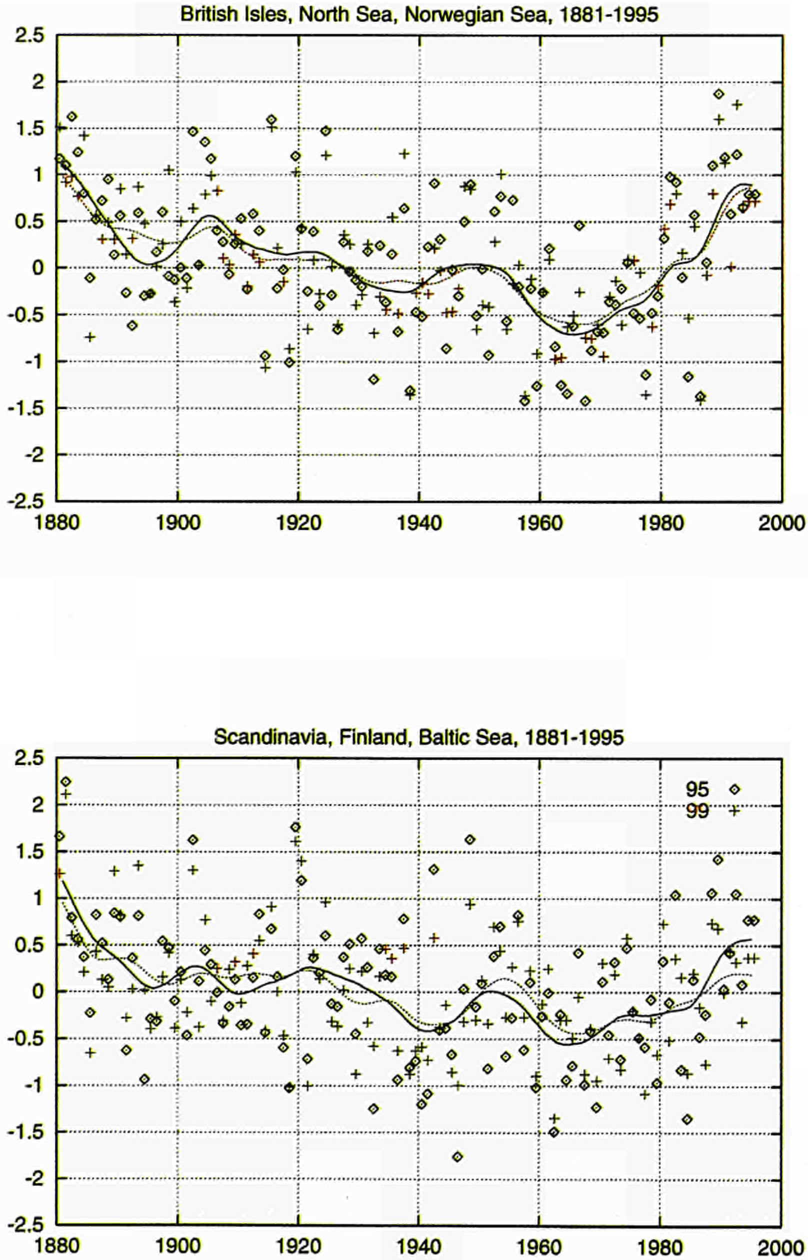

FIGURE 2.9. Annual values of standardised 95- (diamonds) and 99- (crosses) percentiles of geostrophic winds derived from station observations of MSLP using the triangle method in the years 1881-1995. Average of 10 triangles over the British Isles, the North Sea and the Norwegian Sea are plotted in the top panel and 10 triangles over Scandinavia, Finland and Baltic Sea are plotted below. The full and dotted lines show the results of applying a Gaussian filter (with standard deviation of 3 years) for the 5 and 99 percentiles, respectively. (From Alexandersson et al., 1997)

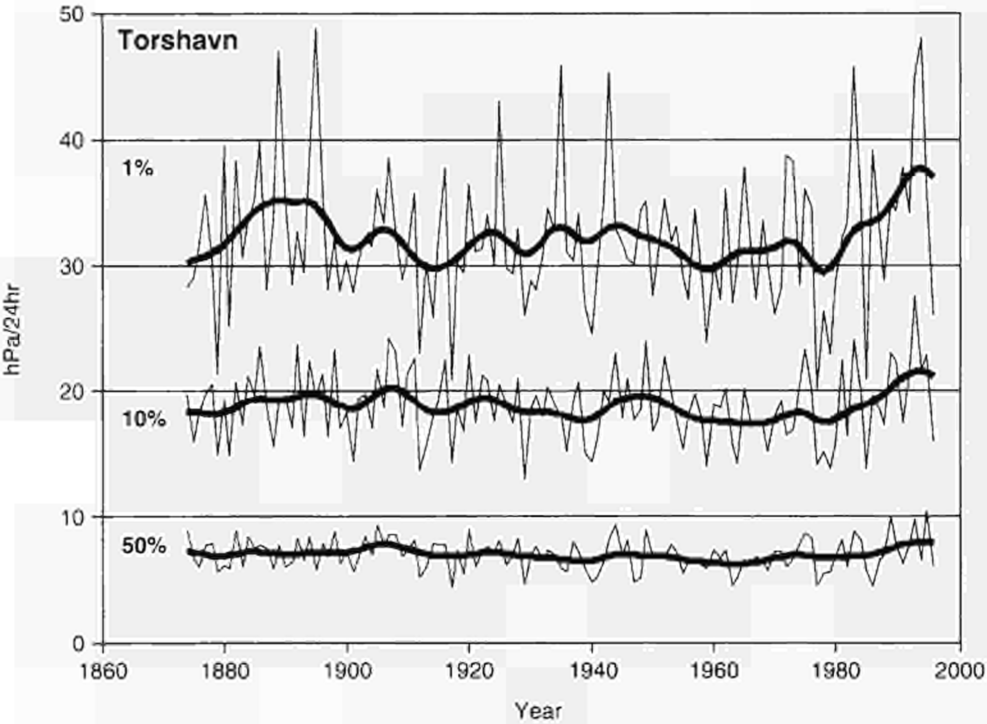

FIGURE 2.10. 1%, 10% and 50% exceedance levels of absolute 24 hour pressure tendency observed 3 times daily for each winter (DJF) for the station Torshavn (62,01N ; 6 46W). The thick curves are obtained by applying a Gaussian filter with a standard deviation of 3 years. (From Schmith et al., 1997)

Another proxy storm data time series is provided by high-frequency sea level variations at tide gauges as suggested by de Ronde (cf. von Storch *et al.*, 1994; von Storch and Reichardt, 1997). Such a time series can be constructed by subtracting the annual mean water level from the data. It is fairly homogeneous because changes in the mean water level mainly reflect processes unrelated to the storm activity, such as local anthropogenic activity (e.g., harbour dredging), mean sea level rise or land sinking. Note, however, that the magnitude of a surge for a given atmospheric forcing may have changed over time due to changes in local/regional bottom topography. After subtraction of the annual mean, the intra-annual distributions of water level variations are left, and intra-annual quartiles are determined as proxies for storm and surge activity as in the case of geostrophic winds and pressure tendencies discussed above. Figure 2.11 shows an example of this for the gauge stations at Den Helder (The Netherlands) and at Esbjerg (Denmark). At Den Helder some corrections have been applied to deal with the effect of closing the Ijsselmmer in the 1930s. The time series of annual exceedance levels show some variations and the two stations are well correlated in time. There is a tendency for an upward trend in the later decades (90% and 97% percentiles).

Summarising, there are some indications that the last decades have been characterised by a higher than normal level of storm and surge activity over parts of the North Atlantic/Northern European region. However, this level does not appear to be significantly higher than what has been seen near the beginning of the present century.

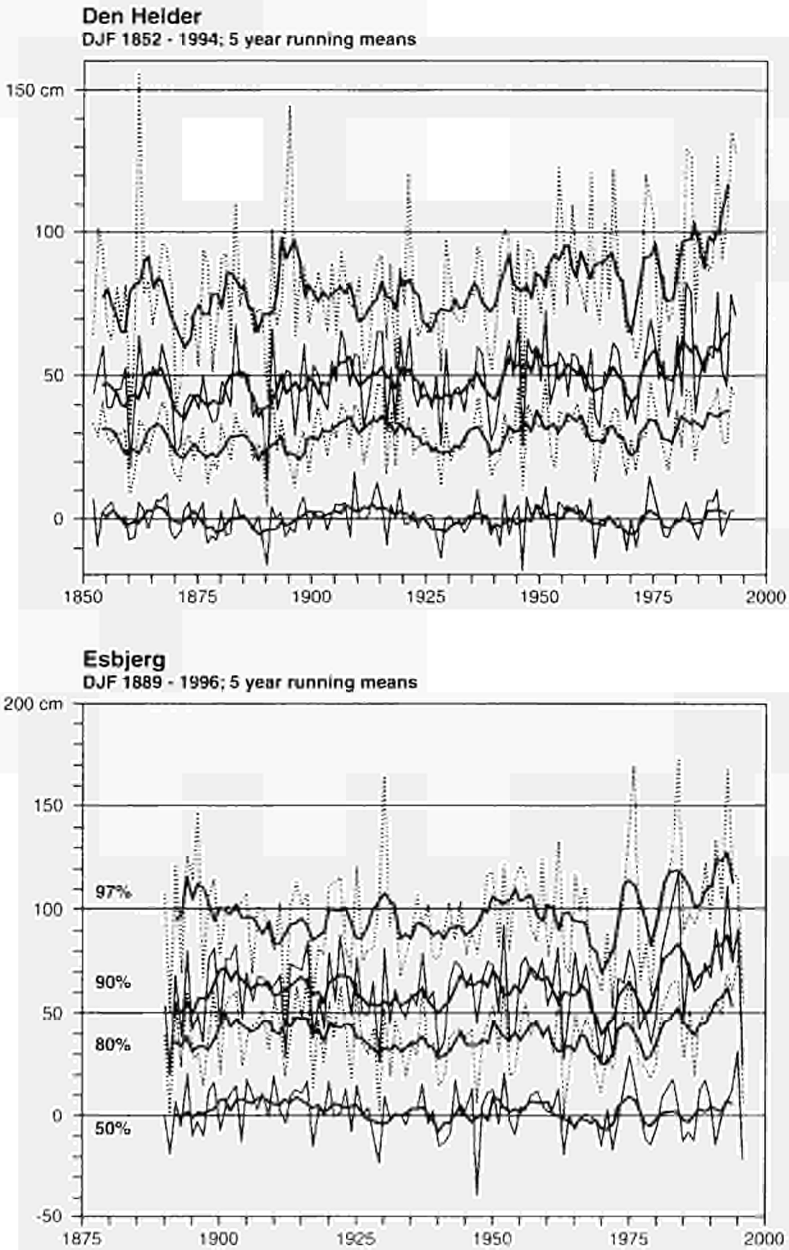

FIGURE 2.11. 50%, 80%, 90% and 97% annual percentiles of high water at the gauge stations at Den Helder (The Netherlands), 1852-1994, and at Esbjerg (Denmark), 1889-1996. Thick curves are 5 year running means.

2.5.1.3 Wave reconstruction in the 20th Century

Data about wave height are available from reports about visual assessments from ships and lighthouses, from wave rider buoys and ship borne wave riders at ocean weather stations; also wave height maps have been constructed for the purpose of ship routing from wind analyses. These data are sparse, and suffer from inhomogeneities of various kinds (cf. WASA, 1994). Analyses of these data have revealed in part a substantial worsening of the wave climate in the North Atlantic (Carter and Draper (1988), Bacon and Carter (1991), Hogben (1994), Bouws *et al.,* (1996)).

To avoid dramatic statements about increases in wave heights which merely reflect observational inhomogeneities, one has to consider more indirect data, as was the case with past storminess. One way to follow is to first identify relationships in modern data between waves and the more homogeneously observed monthly mean sea level pressure field. Then this relationship is used to estimate wave heights in the past where only the atmospheric pressure is observed reasonably well. Such a procedure was followed by Kushnir *et al.* (1995) and is also the approach undertaken in the WASA project.

First a state-of-the-art wave model, WAM (Komen *et al.,* 1994), was integrated with 40 years (1955-94) of wind analyses. Simulations were first carried out over the whole North Atlantic region with a coarse grid wind data set (FNOC) generating boundary conditions for a fine mesh limited area simulation, forced with high-resolution (50 by 50 km) wind analyses prepared by the Norwegian Weather Service DNMI (see Günther *et al.,* 1997). The annual maxima averages and 50, 90 and 99 percentiles were computed in each grid point and each year for both wind and simulated significant wave height[2]. As a representative example Figure 2.12a & b show the linear temporal trend for the 90-percentiles. The figure shows an increase in significant wave height in the north-west approaches of the North Sea. The high positive trends in wind speeds at the western boundary not clearly reflected in wave height may be due to boundary zone problems (differences between coarse and fine mesh wind driving) which again may be related to inhomogeneities in the FNOC and/or DNMI derived wind fields - see discussion in Günther *et al.* (1997).

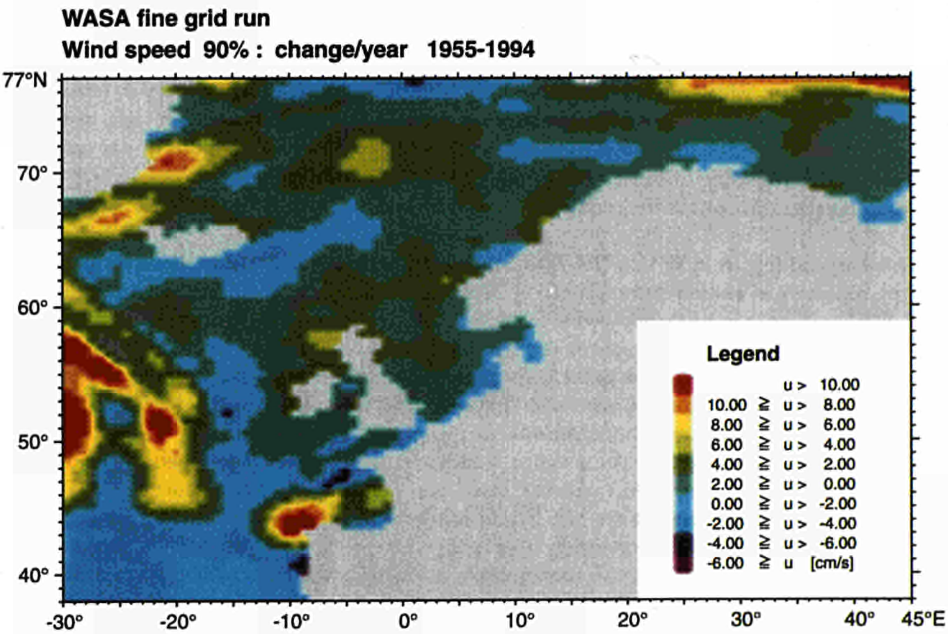

FIGURE 2.12a. Distribution of 40-year trends in 90-percentiles of wind velocity Winds are from the DNMI 50x50 km (+FNOC, locally) analyses driving a WAM model fine mesh simulation.

[2] The significant wave height is proportional to the integral over the energy spectrum.

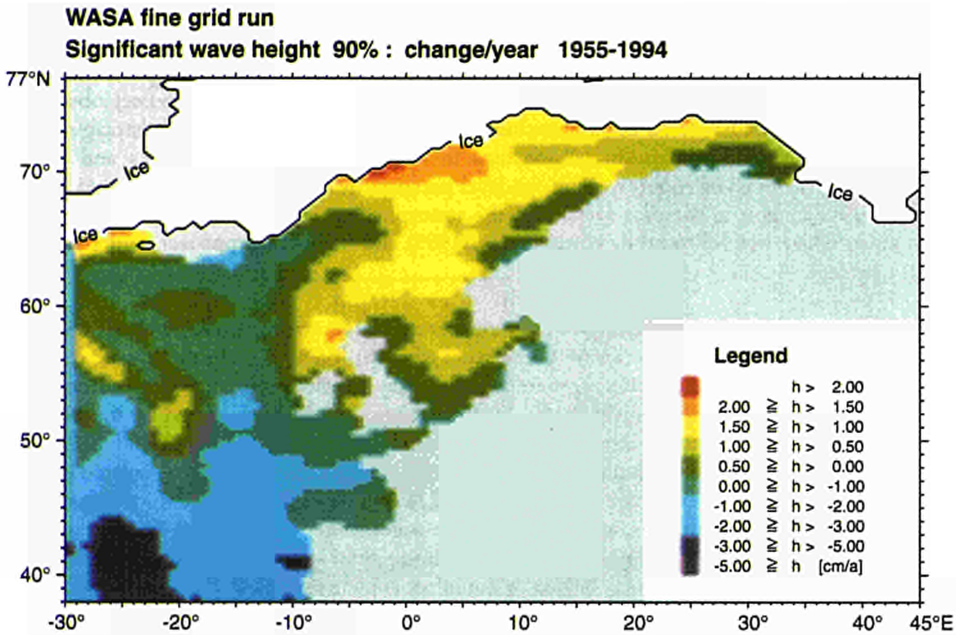

FIGURE 2.12b. Distribution of 40-year trends in 90-percentiles of hindcasted significant wave height.

In the second step an empirical model based on Redundancy Analysis (von Storch and Zwiers, 1998) was built to obtain wave climate further back in time. This model relates intra-monthly percentiles of wave heights at selected locations to gridded monthly mean air-pressure distributions (maps) with the help of the 40 year simulated data; finally observed monthly mean air pressure fields from the beginning of this century were fed into the empirical model to derive best estimates of wave statistics throughout the century[3].

It can be argued (e.g. as in WASA (1997) that the winds driving the WAM model over the entire 40 year period are not entirely homogeneous. This means that also the directly simulated (40 year hindcasted) waves may suffer somewhat from inhomogeneities in the target area south of 70N and east of 20W. However, the predictor data (monthly mean sea level pressure fields) entering the empirical model are considered quite homogeneous[3]. This means that to the extent that the empirical relationships are stationary over time the wave reconstruction made with the empirical model can also be considered homogeneous in the period studied (1899-94)[3]. One of the basic findings using the empirical model is that it identifies a strong relationship between the so called North Atlantic Oscillation, NAO (see e.g. van Loon and Rogers, 1978; Hurrell, 1995) and the patterns of waves, even though the NAO does not explain everything. The last 40 years may be compared with the hindcast data made with WAM, whereas the first five decades represent a best guess and can not be verified at this time. Figure 2.13 shows the 80% and 90% quantiles of wind sea height (height of wind forced surface waves) for both the empirically reconstructed time series 1899-94 and the WAM-hindcasted time series 1955-94 at the location of the oil field "Brent"

[3] Note that the problem with inhomogeneities in the maps of atmospheric sea level pressure is much reduced when monthly mean fields, rather than daily weather maps, are considered. This is because of the greater smoothness of monthly mean fields.

(60N,1.5E)[4]. The results for the swell are similar but smaller and not shown for the sake of brevity. Figure 2.13 shows that for the past four decades, the similarity between hindcast and statistically derived heights is good, and the statistical model confirms the hindcasted increase. However, this increase appears as "normal" when compared to the changes which may have taken place earlier in this century. Indeed, waves as tall as nowadays seem to have taken place in the first two decades, when the NAO was strongest; in the 1920s the NAO weakened significantly (van Loon and Rogers, 1978), and our statistical model indicates that concurrently the height of the waves dropped by several tens of cm/year. In a similar exercise for the Baltic, Mietus and von Storch (1997) found also no systematic worsening of the wave climate in the Baltic.

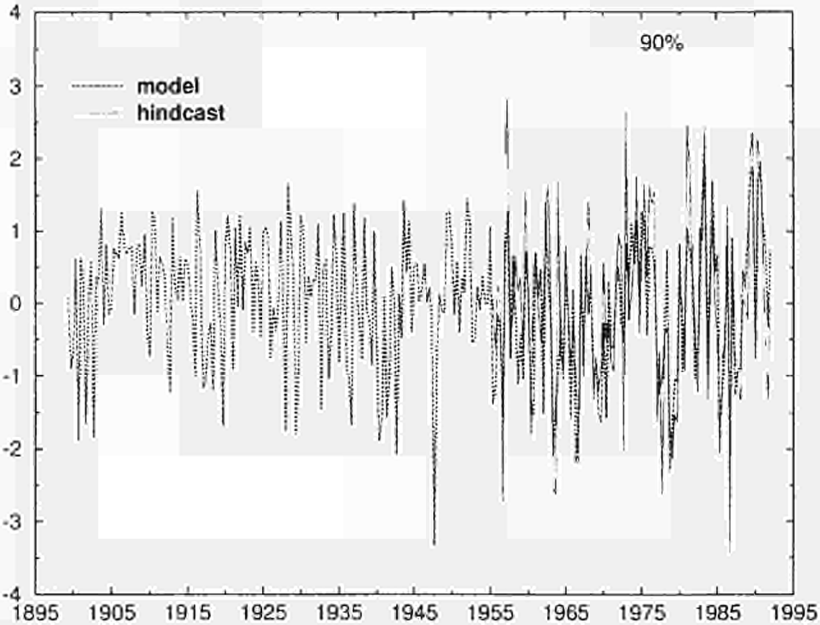

FIGURE 2.13. Time series of intra-monthly 90 percentiles of significant wave height from the hindcast (solid line) and as derived from monthly mean Atlantic air pressure through a statistical model.

2.5.1.4 Future storm, wave and surge climate

Concerning the second main question raised above one has to rely on data from climate models. On the largest atmospheric scales there is a good qualitative agreement between different climate models as to the time-mean temperature greenhouse signals. These signals are (e.g. Hall *et al.* 1994):

- a relatively large warming near the ground at high latitudes,
- a maximum warming in the tropical upper troposphere, and
- a global increase in the water-vapour content.

[4] Measurements of wind wave and swell at Brent in recent years compare well with the hindcast and the reconstruction (Günther *et al.* 1997)

These features of the time-mean state are associated with different changes of the extra-tropical storm tracks with the decreased low level temperature gradient (first signal) corresponding to reduced activity and opposite with the last two signals. Since it is difficult to deduce much at the quantitative level from the long term mean signals one has to study the storms simulated by the climate models, i.e. day-to-day variability. The model data can then be interpreted further by letting them drive wave and surge models.

2.5.1.5 Storms

One may use the output from climate models to estimate likely changes in storm activity in different regions. Due to limitations in mainly the parameterisation of non-resolved physical processes and lack of sufficient spatial resolution even the newest generations of climate models show important systematic errors in both their long term flow and in the variability along the average. For these reasons one must be very cautious interpreting the model output, particularly on regional scales and on fast temporal scales, since e.g. the greenhouse warming signals on these scales are quite different from model to model. There have - nevertheless - been several model studies dealing with analysis of extra-tropical storm activity in a warmer climate, e.g. Hall *et al.* (1994), Lunkeit *et al.* (1996), Beersma *et al.* (1997). Hall *et al.* (1994) analysed the simulated storm tracks in two equilibrium integrations with a medium resolution grid point model. The model was coupled to a mixed layer ocean model and the statistics for 10 years of doubled CO_2 concentrations were compared to 10 years of control conditions. Hall *et al.* (1994) found that the storm tracks are stronger by approximately 10% and shifted downstream in both the Atlantic and the Pacific storm tracks. This intensification is strongest in the Atlantic sector. They also found that the storm tracks were shifted northward by about 5 degrees. Lunkeit *et al.* (1996), using data from a coupled low resolution climate model, find little modification of the Pacific storm track while the Atlantic middle tropospheric storm track shows a downstream intensification with increased high-frequency variability over Central Europe and eastward. Near the surface the signal is much smaller with an increase in storm activity along the north-western European coasts. These findings are somewhat different from those obtained in WASA where data from a so called time slice experiment with the ECHAM3 atmospheric climate model at high (T106) horizontal resolution was used. This set of simulations - each covering only 6 years - show that there are certain differences between the $2xCO_2$ simulation and the corresponding control period (Beersma *et al.*, 1997). These are mainly a weak increase of storm activity over the Bay of Biscay and the North Sea (i.e. downstream enhancement of storm track), while storm action slightly decreases along the Norwegian coast and over most of the North Atlantic. Note, however, that these differences are generally smaller than what can be found by comparing two different randomly selected 6 year periods of observations and that they for this reason can not be considered significant.

2.5.1.6 Waves

The ECHAM3 high resolution winds have been used to drive the wave model, WAM (Komen *et al.*, 1994). Figure 2.14 shows the difference in 90% quartiles of significant wave height between the $2xCO_2$ simulation and the corresponding control period (Rider *et al.*, 1996). The spatial distribution of wave climate changes are in fine agreement with the changes in the driving winds simulated in ECHAM3 (not shown) and show some similarities with the patterns of changes derived in the 40 year historical hindcast - Figure 2.13. It must be emphasised, however, that the differences fall well within the limits of interannual variability observed in the past.

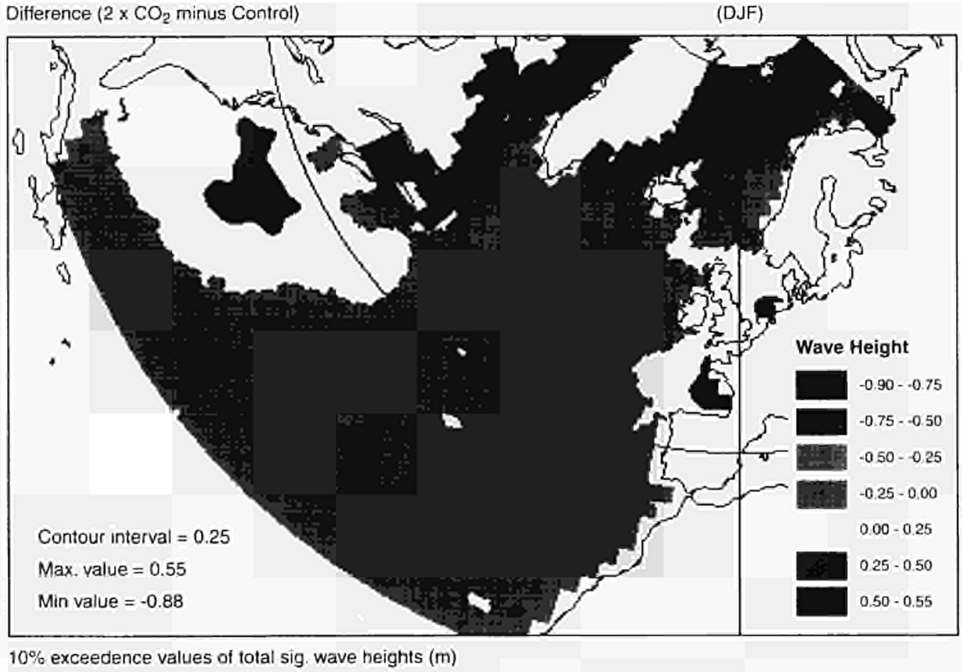

FIGURE 2.14. Difference between the 90 percentile level of total significant wave height in 6 winters (DJF) of CO_2 forced climate and 6 control climate winters as simulated by a wave model (WAM). WAM was driven by atmospheric conditions taken from two time slice simulations with a high resolution global atmospheric climate model (ECHAM3/T106).

2.5.1.7 Future surges

The climate model data has also been used to drive a surge model covering the northern part of the European shelf (Flather and Smith, 1997). From these simulations it can be concluded that due to a shift in the steering wind direction in the scenario there is a general increase in surge levels, in particular more north-westerly storms hit the North Sea area leading to more severe surge events in the area of the German Bight, see Figure 2.15. One must, however, note again that the T106 simulations used to drive the surge model only cover a short period and that the high scenario values may solely be due to under-sampling. It should be mentioned that the surge model has been verified successfully in a 40 year hindcast simulation where it was forced with the same observed gridded wind/pressure data driving the previously mentioned wave model hindcasts (Flather *et al.*, 1997).

The effect of global warming on storm surges in the North Sea area was also studied with a statistical model (von Storch and Reichardt, 1997). The statistical model was based on a link between intra-monthly percentiles of storm-related high tides and the monthly mean air pressure distribution (similar to the 2nd step in the analysis of waves in the present century discussed above). When applied to T106 data, a slight increase of one or two decimetres was calculated for the 90 percentiles for the time of doubled carbon dioxide in the atmosphere. When an alternative global warming scenario was used, prepared by a 30 year time slice simulation with a T42 model, the changes were found to be almost nil.

FIGURE 2.15. Difference between distribution of the 5 year return surge elevation (cm) in 5 years of CO_2 forced climate and 5 control climate years as simulated by a surge model. The surge model was driven by atmospheric conditions taken from two time slice simulations with a high resolution global atmospheric climate model (ECHAM3/ T106).

2.5.1.8 Wind and pressure forcing and the implications for sea level

Eigil Kaas and H. von Storch

As with short term sea level variations and extreme events (section 2.5.1), variations on time scales from weeks to decades or even longer are closely related to the internal dynamics of the atmosphere/ocean system. In the analysis of possible trends in observed globally averaged sea level, the high-frequency ocean signals, i.e. waves, swells, tides and surges are relatively easy to remove by filtering techniques, as illustrated by e.g. Sturges (1987). This is, however, not the case with low-frequency variability and the lowest frequencies that can generally be resolved with longer available sea level records are of the same amplitude as (or larger than) the globally rising sea level signal (IPCC, 1995).

On the inter-annual to decadal periods, sea level fluctuations are often driven by atmospheric wind and pressure forcing. Slow variations in wind stress lead to a pile up of water which is modulated mainly by the rotation of the Earth, friction and the local topography. The atmospheric pressure has an inverse effect on sea level. The direct barometric effect - not considering so called geostrophic adjustment - of a 1 hPa decrease in atmospheric pressure is a 1 cm increase in sea level. For several phenomena a coupling between atmosphere and ocean takes place. It should be noted that it is not only the mechanical effect of wind/pressure forcing that is important. It is also the thermodynamic processes which must be considered in interpreting long-term sea level variations. Of particular importance - in connection with climate change - are the differential oceanic heating and changes in salinity. As an example, changes in the thermohaline circulation and the production rate of North Atlantic Deep Water (NADW) may change the northward transport of upper ocean warm water, which could lead to an enhancement of sea level rise in the mid-latitudes of the North Atlantic (Mikolajewicz *et al.* 1990).

2.5.1.9 Phenomena related to low but frequent sea level variations

In the Tropical Pacific - on the "short-term inter-annual" time scale - the El Niño-Southern Oscillation coupled atmosphere/ocean phenomenon can clearly be observed by in-situ tide gauges and satellite altimetry. During the positive phase sea level is lower than normal at the western Tropical Pacific and during the negative phase (La Niña) it is high. The opposite situation dominates the eastern Tropical Pacific - see Wyrtki (1979, 1985), Miller and Cheney (1990) for more details.

Less well known are the 5-8 year variations seen for gauges at San Francisco and Honolulu which are related to oceanic baroclinic Rossby wave propagation (Sturges, 1987).

Long homogeneous atmospheric and oceanic observational records which can be used to study air-sea interactions, including sea level, are unfortunately very sparse or non-existing. In recent years where coupled atmosphere-ocean models have become available it has, however, been possible to study many processes and events in more detail and even discover new phenomena. An example is the air-sea interactions between the sub-tropical gyre circulation in the North Pacific and the Aleutian low pressure system which has been found in a version of the coupled AOGCM at the Max Planck Institute for Meteorology (MPI) in Hamburg (Latif and Barnett, 1994). The sea level variations associated with this inter-decadal mode are of the order 3 - 4 cm.

2.5.1.10 Sea level variations in the European region

In recent years there has been much focus on the behaviour of the thermohaline circulation in the North Atlantic, the NADW formation and the Conveyor Belt Circulation, and the possible existence of multiple equilibrium states for the global ocean circulation (e.g. Rahmstorf, 1995). These are processes, related to anthropogenic/natural climate change, which would result in regional changes in sea level that are larger than the globally averaged change, particularly in the North Atlantic Region (Mikolajewicz *et al.,* 1990; Cubasch *et al.* 1994).

In addition to these long term anthropogenic (or natural) potential threads, European coastal sea level variations are strongly influenced by the direct atmospheric forcing. In a recent paper by Heyen *et al.* (1996) it was thus shown that Baltic sea level variations in winter in the present century can be modelled quite well in a linear regression model using only the varying spatial distribution of monthly mean sea level air-pressure (MSLP) over the North Atlantic as input. This so-called down-scaling model is based on canonical correlation analysis (CCA) which identifies mutual anomaly patterns of MSLP and sea level. Based on these patterns the empirical model can be used to estimate sea level in months were only MSLP is available and it can be used to interpret the climate model simulated MSLP in terms of sea level. The model was trained (built) on data for the winter months 1951-70. The first pair of patterns is shown in Figure 2.16. Figure 2.17 depicts the observed and estimated sea level anomalies for the entire period 1899-1987. It is seen that there are high correlations between estimated and observed values, particularly in the northern part of the Baltic. Heyen *et al.* (1996) argued that it is wind stress (which can be related to MSLP) which forces the variations in the Baltic sea level. Note, that long term trends were excluded from this historical analysis. The down-scaling model was next used to interpret the MSLP patterns simulated in a transient coupled greenhouse gas sensitivity experiment, Cubasch *et al.* (1992), with a former version of the MPI model (ECHAM1). The result was a weak net upward trend of 0.01 cm/year in the simulated Baltic sea level which was masked by decadal variability. Note, however, that this result is very dependent on the GCM data used as input.

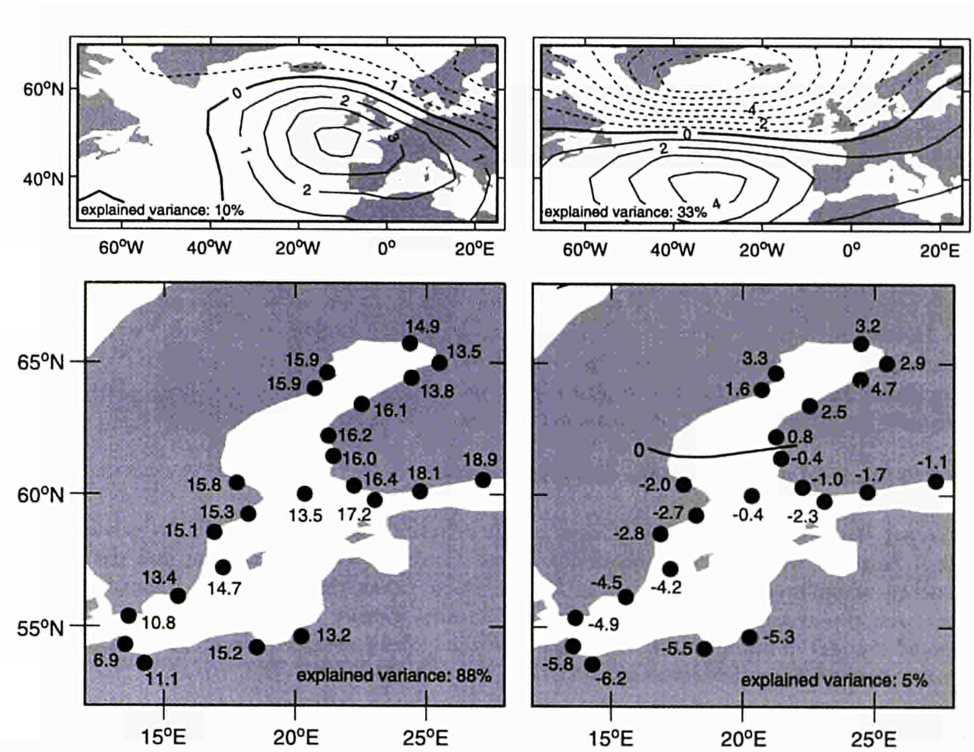

FIGURE 2.16. The first pair of CCA patterns depicting anomalies in hPa and cm that belong together, i.e. in months where the observed anomalies in MSLP are similar to the upper panel the sea level anomalies are close to those in the lower panel. The time series associated with the first pair are correlated with 8. The amplitudes on the plot are defined in such a way that the CCA-timeseries possess mean 0 and standard deviation 1. Hence the values in the CCA-patterns can be interpreted as "typical" anomalies in hPa and cm.

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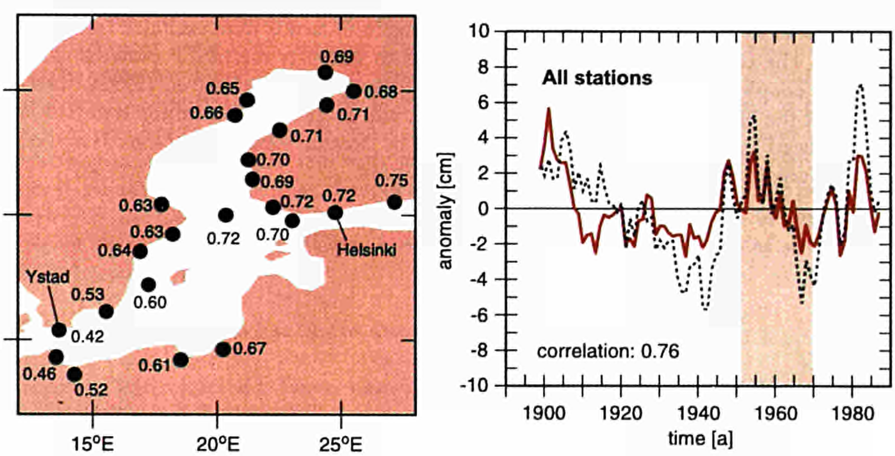

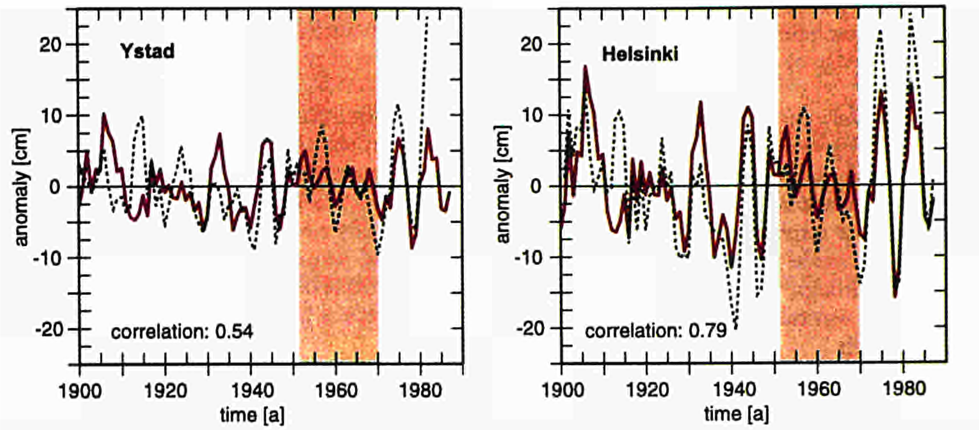

FIGURE 2.17. Detrended winter air-pressure observations 1899-1987 have been downscaled to derive sea level anomalies. These are compared with detrended sea level observations: The map (previous page) depicts the local correlation between estimated and observed monthly means. The graphs show estimated (solid line) and observed (dashed line) anomalies for all stations (overleaf, right) and for two selected stations, Ystad (left) and Helsinki (right). The graph data have all been smoothed by a 3 year running mean. The shaded area indicate the training period 1951-70.

2.6 CRUSTAL MODELS

Hans-Peter Plag

In many observations of global change phenomena including relative sea level changes and changes in ice mass, crustal movements due to tectonic, anthropogenic or other causes are masking the global change signal that is being sought. Therefore, GPS or other space-geodetic observations as well as absolute gravity measurements are required particularly at tide gauges but also at other locations in order to get a clear picture of crustal motion. It should be pointed out that combined GPS and absolute gravity observations give valuable information on the underlying mechanisms. Thus, these observations provide constraints necessary for the validation of global and regional geophysical models of present-day crustal movements. Such models are required for decontamination of observations in order to separate relevant global change signals from other disturbing influences.

In setting up such models the problem of a reference system suitable for relating observations and modelling results as well as providing the required long-term stability deserves particular attention. Recent progress in space-geodetic techniques and world-wide co-ordination of permanent observations particularly with GPS (within the International GPS Service for Geodynamics, IGS) have laid the foundation for the realisation of the required high-accuracy reference system. For achieving a crustal movement model as well as for the separation of the various global change phenomena contributing to crustal motion, the continuation and constant improvement of the reference system is a key issue.

The temporal variability of the geoid due to mass re-locations (e.g. melting or accretion of ice sheets) and elastic deformations of the Earth induced by mass movements are, on the one hand, affecting the distribution of the ocean water significantly, particularly if additional water is added to the ocean from the ice sheets. Thus, a melting of a layer of the Greenland ice sheet will be associated with a distinct geographical pattern in sea level rather different from the pattern associated with a similar melting of a layer from the Antarctic ice sheet. On the other hand, temporal changes of the geoid are indicative of mass movements within the Earth systems. Therefore, crustal movement models should be associated with geoidal models of high spatial and temporal resolutions. Dedicated space-born gravity missions such as GRACE will greatly improve our knowledge of the geoid and allow for such a combination of geoid models with crustal models.

Unfortunately, the currently available global crustal movement models such as NUVEL1A (DeMets *et al.,* 1994) give only the horizontal velocity of the main tectonic units. No global model exists for the vertical component. It can be expected that the observations of the global space-geodetic networks eventually result into such a model. However, both the horizontal and vertical components display considerable small-scale spatial variability necessitating the augmentation of any global model by regional to local models. The regional densification particularly of the IGS network through, for example, the EUREF network, will provide observations to constrain regional models of vertical crustal movement. Unfortunately, very little is being done currently to establish such models.

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3

EXTREME COASTAL FLOODING

DAVID E. SMITH, *Chapter Editor*

CESAR ANDRADE, ALASTAIR DAWSON, CALLUM FIRTH, CONCEÇÃO FREITAS, & SHAOZHONG SHI & DAVID E. SMITH, *Chapter authors*

9 September 1840: Shetland Isles

"One of the most severe gale of wind known in the memory of the oldest in Shetland at this season of the year. The wind came out of the north north west and was of hurricane force. Five boats went with all hands that night, and over 30 men lost their lives. In Lerwick there was other chaos, boats dragging and driving everywhere, many being smashed to pieces on the shore. For 15 years afterwards folk still spoke of the Hungry Forties. Boats which had been hauled ashore in 1840 were never launched again, were left to moulder and rot on their shores."

(Anon., quoted in Henderson, 1980)

3.1 INTRODUCTION

David E. Smith

The impact of storms and flooding on the coastal zone is frequently severe, and can involve great loss of life and extensive damage to property. Its importance in the public consciousness throughout history is attested to in the many accounts, such as the one, above which in greater or lesser degree remark upon both their social and economic consequences.

With increasing recognition that there is a need to determine the origin and impact of coastal flooding, recent years have seen the advent of a number of concerted research programmes in this field. Thus, global programmes such as the UNESCO-supported International Geological Correlation Project (IGCP) 367 *"Rapid Coastal Changes"*, and a number of EU projects (notably EV4C-0047, *The spatial and temporal variability of major floods around European coasts* (1987-1990); EPOC-CT90-0015, *Climate change, sea level rise and associated impacts in Europe* (1990-1992); EV5V-CT93-0266, *Relative sea level changes and extreme flooding events around European coasts* (1993-1995); EV5V-CT93-0258, *The impacts of climate change and relative sea level rise on the environmental resources of European coasts* (1993-1995); EV5V-CT94-0445, *Climate change and caostal evolution in Europe* (1994-1996); EV5V-CT92-0175, *The genesis and impact of tsunamis on European coasts* (1992-1995); ENV4-CT96-0297, *The genesis and impact of tsunamis on European coasts: tsunami warning and observations* (1995-1998); EV5V-CT94-0503, *Variability of the North Atlantic Storm Track* (1995-1996); EV5V-CT94-0506, *Impact of storms on waves and surges: changing climate in the past 100 years and perspectives for the future* (1994-1995); and ENV4-CT97-0488, *Storminess and environmentally sensitive Atlantic coastal areas of the European Union* (1997-2000)) as well as the work of the INQUA Shorelines Commission and numerous national projects, have addressed this problem, demonstrating the importance which the subject has assumed both in the minds of scientists and for the larger community. With over 50% of the world's population living in coastal areas (UNEP, 1990), coastal storminess possibly increasing (e.g. Raper, 1993), and widespread rises in relative sea level bringing ever greater areas of the coastal margin within reach of flooding from the sea, such interest in the subject is understandable.

This chapter does not however attempt a comprehensive examination of recent work in Europe on coastal flooding. The major factor of flooding by rivers is not considered nor are the consequences of flooding for terrestrial (i.e. non-marine) processes or for the groundwater reservoir. It is concerned with empirical studies of coastal flooding and of the effects of such flooding on coastal marine processes, complementing the modelling approaches addressed in chapter 2, above. It considers coastal flooding on timescales of minutes to 1000 years, focusing upon European coastlines. The chapter will first consider developments in studies of the processes by which coastal flooding may affect coastal geomorphology, following which European examples of flooding due to secular change in sea surface levels, storm surges and tsunamis, will be examined. The locations referred to are shown in Figure 3.1.

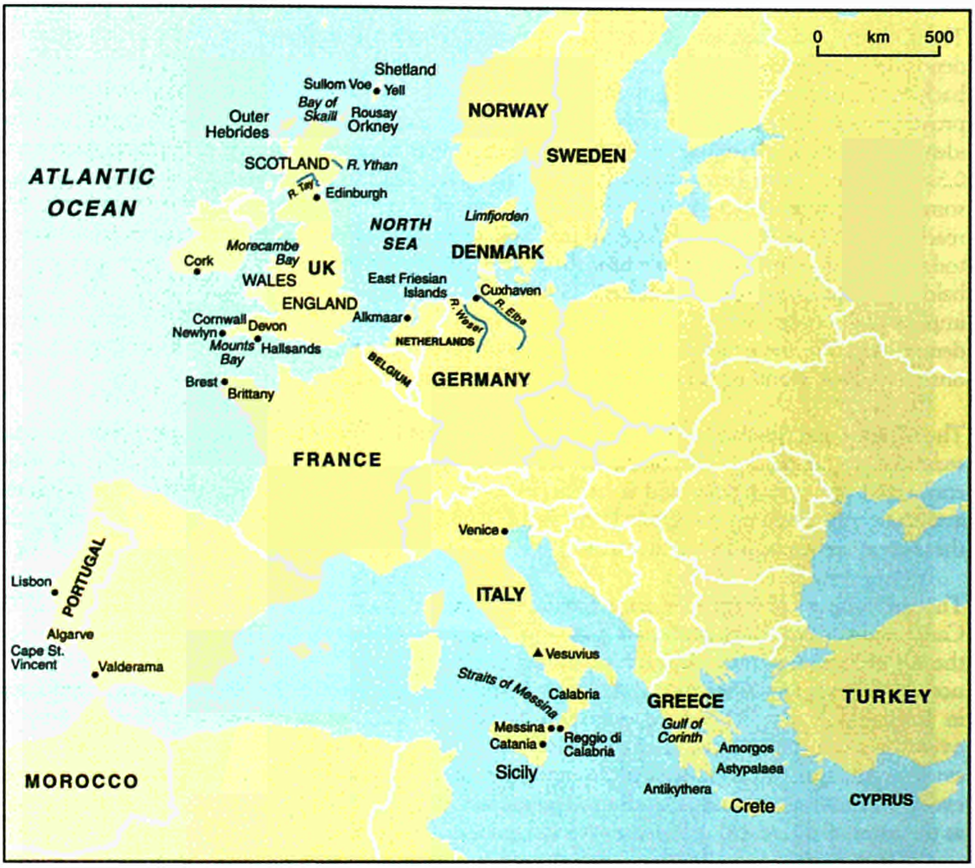


FIGURE 3.1 *European locations discussed in this chapter.*

3.2 THE EFFECTS OF MARINE FLOODING ON COASTAL PROCESSES

David E. Smith

The processes by which flooding from the sea has influenced the physical development of the European coastline have taken place against the background of a number of external factors. Locally, these include offshore bathymetry, coastal configuration, exposure, geology, sediment supply, tide and current activity. More widely, changes in the level of the surrounding sea surface are fundamental, although broad tectonic effects, notably isostatic changes, and localised neotectonic effects, as well as geophysical and other changes have modified the effect of sea surface changes, most authors preferring to use the term **relative** sea level in recognition of these effects.

Along the European coastline, the power of storm waves has been attested to in a number of studies in the past (see examples in King, 1973). Recent studies in Europe have further emphasised this. For example, along the exposed western coastline of the Orkney and Shetland Island groups, U.K., cliff-tops up to at least 30m above Ordnance Datum Newlyn (OD) (the UK levelling datum) are often garlanded with accumulations of gravel and boulder-sized angular material, sometimes imbricated, (Figure 3.2). Hall (1996) has maintained that such material has been thrown up by storm wave activity. On the island of Rousay, Orkney, he claims (pers. comm.) that there may be evidence of discrete accumulations, indicating evidence of a number of storms, raising the interesting possibility of identifying local sequences of events.

The Orkney and Shetland Islands also contain evidence that storms may have resulted in the deposition locally of sheets of sand. These islands are widely covered in blanket peat, often dating back to the early Holocene (e.g. Birnie, 1981). At some coastal locations, layers of sand of marine provenance have been traced within the peat. At Sullom Voe, Mainland, Shetland, Birnie (*ibid*) identified one such layer, up to 30cm thick (Figure 3.3), tapering inland over a distance of up to 0.5km and containing brackish/marine diatoms. Later, this layer was dated as having accumulated sometime between 5315$\pm$45 and 5700$\pm$45 ^{14}C years BP (Smith, 1993). The Sullom Voe layer reaches to at least +9.0m OD, where the high water mark of ordinary spring tides (HWMOST) today is below +3m OD. At the time the layer accumulated, relative sea level lay perhaps over 5m below present according to Mörner's (1980) NW Europe sea level curve, and the altitude thus implies high energy events. Tsunami origin is possible, although no tsunamis have so far been demonstrated in the general area for the time involved. Storm surge origin, probably assisted by onshore winds, seems equally possible, and further work is awaited to determine its origin.

The Orkney and Shetland evidence indicates that extreme storms may have the power to deposit sediment to altitudes substantially above contemporary HWMOST. However, the deposits concerned are very patchy and it seems likely that notwithstanding the influence of sediment availability, the flooding was generally localised. The erosional and depositional effects of storms in these areas are likely to be highly variable.

The response of barriers to storms has been discussed widely (e.g. Orford *et al.*, 1991 & 1995; Carter and Orford, 1993; McKenna *et al.*, 1993; Regnauld *et al.*, 1996). Rossiter (1954) remarked on the destruction of part of Orford Ness during the 1953 North Sea storm surge, whilst Steers (1953) noted washover fans along nearby Blakeney Point during the same event. A number of studies (e.g. in Regnauld *et al.*, 1996) have examined storm deposits, resulting from barrier breaching or overtopping during storms, now concealed within lagoonal sediments landward of barriers. It is evident that storms can have a profound effect upon barrier development although for a given event, barrier response may vary due to factors such as barrier architecture and composition as well as the external influences outlined above and, increasingly, anthropogenic effects.

The response of beaches to storms has been discussed at length in the literature. Whilst storm activity may generate sediment, storm waves may also comb sediment seawards, thus steepening the beach profile and rendering the coastline more susceptible to erosion. The consequences of removal of sediment from the beach have been illustrated by the example at Hallsands, Devon, U.K. (Tanner and Walsh, 1984) where anthropogenic activity rendered the coastline vulnerable to considerable erosion with severe consequences for the local community.

A major cause of flooding is tsunami activity, where a growing literature is developing on associated effects (e.g. Dawson *et al.*, 1995, 1996; Dawson, 1994; Shi, 1995; Shi *et al.*, 1993). These sometimes major events may achieve both widespread erosion of the coastline and extensive deposition, affecting subsequent coastal development, as Andrade and Freitas discuss below. In Scotland, UK,

FIGURE 3.2 Cliff-top storm deposits, Rousay, Orkney, UK.

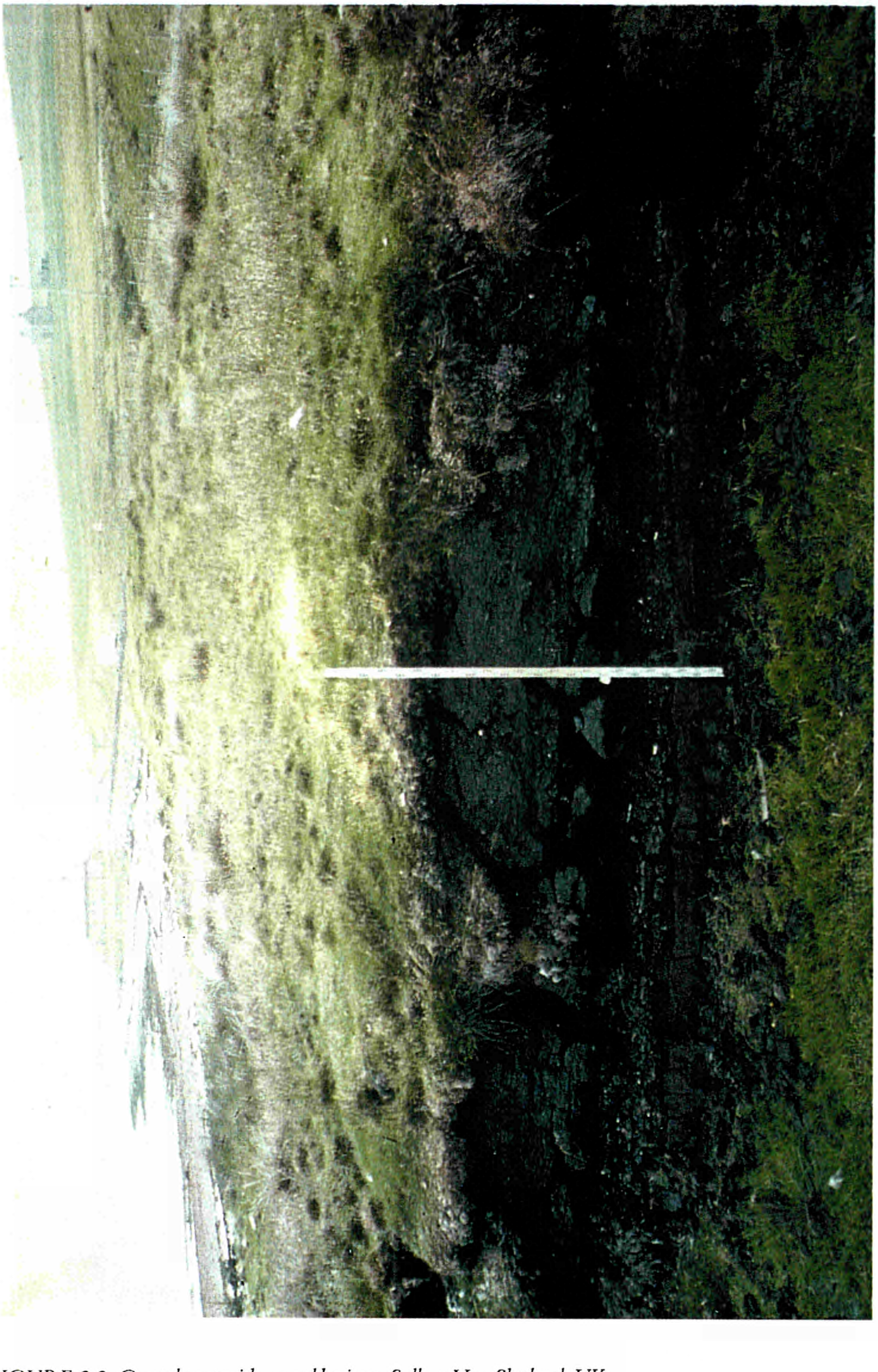


FIGURE 3.3 Coastal peat with a sand horizon, Sullom Voe, Shetland, UK.

the tsunami generated by the Second Storegga Slide of *circa* 7100 ^{14}C years BP (Figure 3.4) (discussed below) caused widespread destruction (Dawson *et al.*, 1990) and locally marked coastal erosion (Firth *et al.*, 1995), although its main feature was the widespread deposition of sand amongst accumulating estuarine silts along the entire eastern coast of mainland Scotland. Other studies attesting to tsunami deposits as well as erosion include in Spain, where the 1755 Lisbon tsunami caused both substantial erosion and deposition along Valderama spit (Dabrio *et al.*, 1998) and in Greece, where high-level gravels on Astypalaea island are attributed to the 1956 southern Aegean tsunami (Dominey-Howes, 1996b). These and other recent studies both in Europe and elsewhere demonstrate that tsunami activity may strongly influence coastal areas, modifying the effects of climate-related coastal flooding and other processes, as the following detailed example illustrates.

3.2.1 An example of rapid coastal change associated with an extreme event: the Algarve coast of Portugal

Cesar Andrade & Conceçáo Freitas

3.2.1.1 Introduction

Any long-term process of change may be considered to be the result of a continuous sequence of multiple micro- to mesoscale events. In this context, the intensity (and effect) of any individual event must be carefully evaluated. In fact, only a (possibly small) fraction of the associated coastal response would contribute to a steady and established long-lasting tendency. The remaining part would be absorbed by the self re-organizing ability of the coastal system. In other words, the response of the coast to each short-term event can be filtered out by integrating it within the longer-term trend of coastal change. This of course applies if the coast is in a state of stable equilibrium with recovery determined by an efficient self re-organizing ability (Figure 3.5), in a steady-state equilibrium with no net change in equilibrium levels, or in a state of dynamic equilibrium with a long-term trend (cf. Butzer, 1982, Bell and Walker, 1994 for a more extensive discussion on the scope and implications of these terms). However, other coasts may be more adequately described as brittle systems, characterised by an evolution which is composed of a succession of metastable equilibria, punctuated by microscale events of rapid change. These events happen when critical environmental thresholds of equilibria are exceeded and induce desproportionate and irreversible responses of the coast (Figure 3.5D).

Research undertaken on the Portuguese coast shows an evolutionary pattern in which episodes of dynamic or steady state equilibrium may be recognised in the geological record of this coast, separated by periodic or aperiodic short lived catastrophic changes after which the coast searches for and reaches a new state of equilibrium. This contribution will present examples and discuss the relevance, implications and signatures of rapid changes recognised as such in the late Holocene sedimentary record of the Portuguese coast. Rapid changes have been identified in the geological record using a suite of environmental indicators, including sedimentological and palaeoecological proxies. Their environmental interpretation relies heavily on the comparison with present-day sedimentary environments.

Rapid changes are identified in this area in terms of: 1) High-energy coastal flooding events and 2) Changes in the environmental setting affecting the sediment budget.

3.2.1.1 High-energy coastal flooding in Portugal

The documentary record demonstrates that the most destructive high-energy events of coastal flooding that affected the Portuguese coast in the last millennium are related to seismic activity of the ocean floor. In particular, the vertical displacement generated by the shallow seismic event of the 1st November AD 1755 generated a large magnitude tsunami that flooded the entire western and southern Portuguese coast as well as the Spanish southern littoral as far as Gibraltar. This tsunami left a varied suite of geological signatures along the Algarve coast and was the most important driving factor of coastal change in that area in the last 500- 1000 years.

FIGURE 3.4 Tsunami deposit consisting of a sand layer (base marked by the red-handled knife point) within estuarine deposits, Montrose, UK.

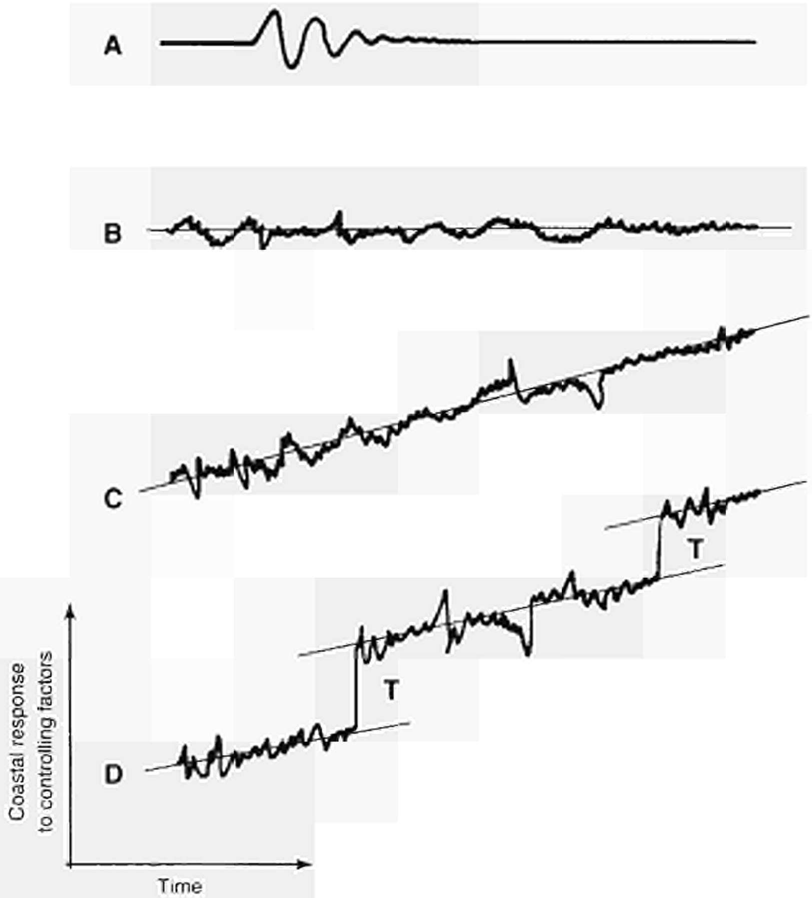


FIGURE 3.5 Schematic representation of different models of coastal change.
A - Stable equilibrium: coastal resilience absorbs the effects of a particular disturbance.
B - Steady-state equilibrium with no long-term trend.
C - Dynamic equilibrium with long-term trend.
D - Brittle system: succession of punctuated equilibria.
Dynamic equilibria with long-term trends separated by thresholds (T).

Andrade (1991, 1992) and Andrade *et al.* (1994) studied the geological signatures of the 1755 tsunami on the barriers of Ria Formosa, a barrier island and lagoonal system that forms the dominant physiographic unit of the central and eastern Algarve coast (Figure 3.6a & b). The interpretation of the surface geomorphology and lithostratigraphy of the barriers allowed the recognition of a catastrophic coastal response to tsunamigenic flooding, followed by a rapid rearrangement of the coastal geomorphological setting. This is summarised in the following model (Figure 3.6a A to D):

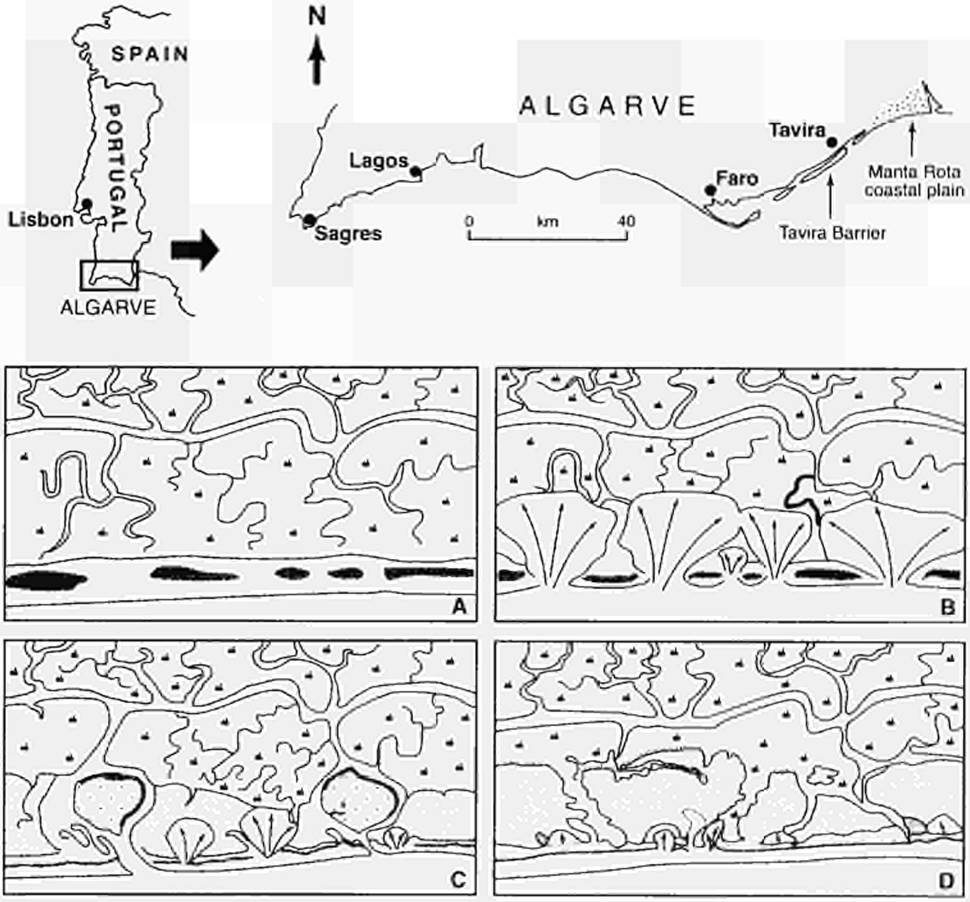


FIGURE 3.6a Location maps of the Ria Formosa barrier system and Manta Rota coastal plain in the Algarve (southern Portugal). A - Pre-tsunami situation. B - Tsunamigenic overwash. C - Development and drifting of short-lived inlets. D - Barrier healed, present-day morphology established (see text or Andrade, 1992 for a more detailed explanation)

A - Pre-tsunami situation. A barrier armed with a low-lying discontinuous foredune ridge lies seaward of an intertidal marsh.

B - The tsunami hits and disrupts the morphological framework of the barrier. A massive overwash of water and sand covers the seaward lagoonal surface, building an elevated platform made of multiple coalescent washovers. A series of small-scale and ephemeral tidal inlets is created. The easternmost islands are completely drowned and destroyed.

C- In the following 15-20 years the newly formed tidal inlets silt up naturally, after limited longshore drifting. The former inlet channels become static washover locations. In the lagoon the tidal drainage network adapts to the imposed rise in the topography. Further east sand is pushed towards the shore and initiates the building up of a new ridge and swale intertidal surface.

D - After complete reconstruction of the barrier, a second generation of beach-foredune structure develops, isolating the former tidal structures that are now incorporated in the backbarrier surface and are reworked by wind and spring-tidal currents. Further east the coastal plain of Manta Rota consolidates and defines the present day coastline

This model is similar in processes and effects to the changes which took place on the NE Japanese coast after tsunamigenic flooding in 1983 (Minoura & Nakaya 1991).

Figure 3.6b illustrates the lithostratigraphic framework of Tavira island. The deposit associated with this episode of rapid change (A) is clearly noticeable in the lithostratigraphical column and marks the exceeding of a threshold induced by one single flooding event.

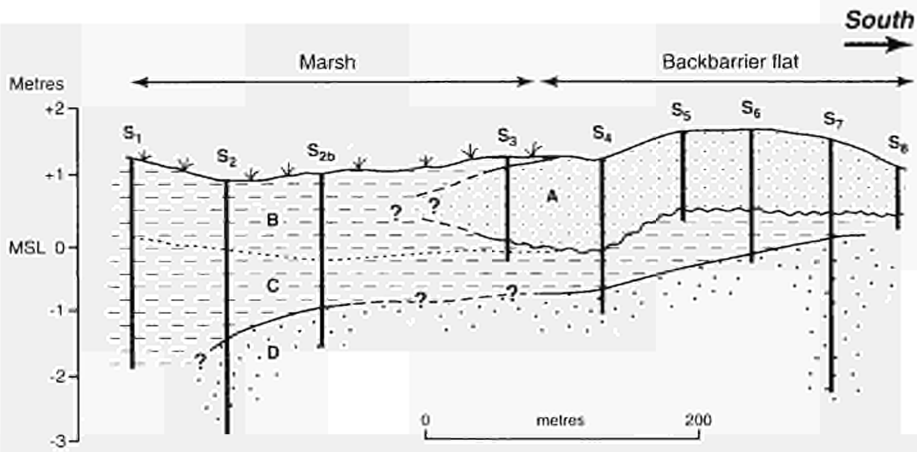


*FIGURE 3.6b North-South cross-section of the back-barrier and marsh of Tavira island (***A*** - Tsunami and tidal delta sands;* **B C** *- Marsh and mud flat sediments;* **D** *- Basal sands).*

In the following 250 years only minor changes of the coastal morphology have taken place in response to small disturbances in the sand budget. Coastal resilience efficiently absorbed disturbances induced by storms or by inlet activity. Further east, between Tavira and the Guadiana river, the tsunamigenic flooding completely drowned and obliterated the barrier islands there, and these never reformed again. The sand was pushed landwards and covered the former lagoonal space, building up the present-day coastal plain of Manta Rota.

3.2.1.2 Changes in the environmental setting of the areas

The reconstruction of the Middle and Late Holocene evolutionary pattern of the Melides and Albufeira coastal lagoons, located in the Portuguese western coast, indicate that both coastal structures evolved through a succession of punctuated equilibria. Long episodes of coastal metastability may be recognised in the sedimentary record, separated by short-term episodes of rapid coastal change, triggered by environmental controls others than eustasy. This conclusion is supported by a thorough analysis of the sedimentary and palaeoecological record; the environmental interpretation presented here relies heavily on comparison with present-day processes and environments and departs from more traditional approaches to palaeoenvironmental reconstruction that tend to ascribe every ecological and sedimentological response of the sedimentary record to a modification of the sea level rise rate or (temporary) reversal of the sea level change signal.

Melides. The coastal lagoon of Melides is a small barrier structure located some 80km southwest of Lisbon and corresponds to the present-day remnant of a valley system drowned by the Holocene transgression. The lagoon is separated from the sea by a reflective sand barrier from which nine short cores were taken for palaeoecological and sedimentological studies (Figure 3.7). Results from these studies may be found in Freitas *et al.* (1993), Queiroz and Mateus (1994).

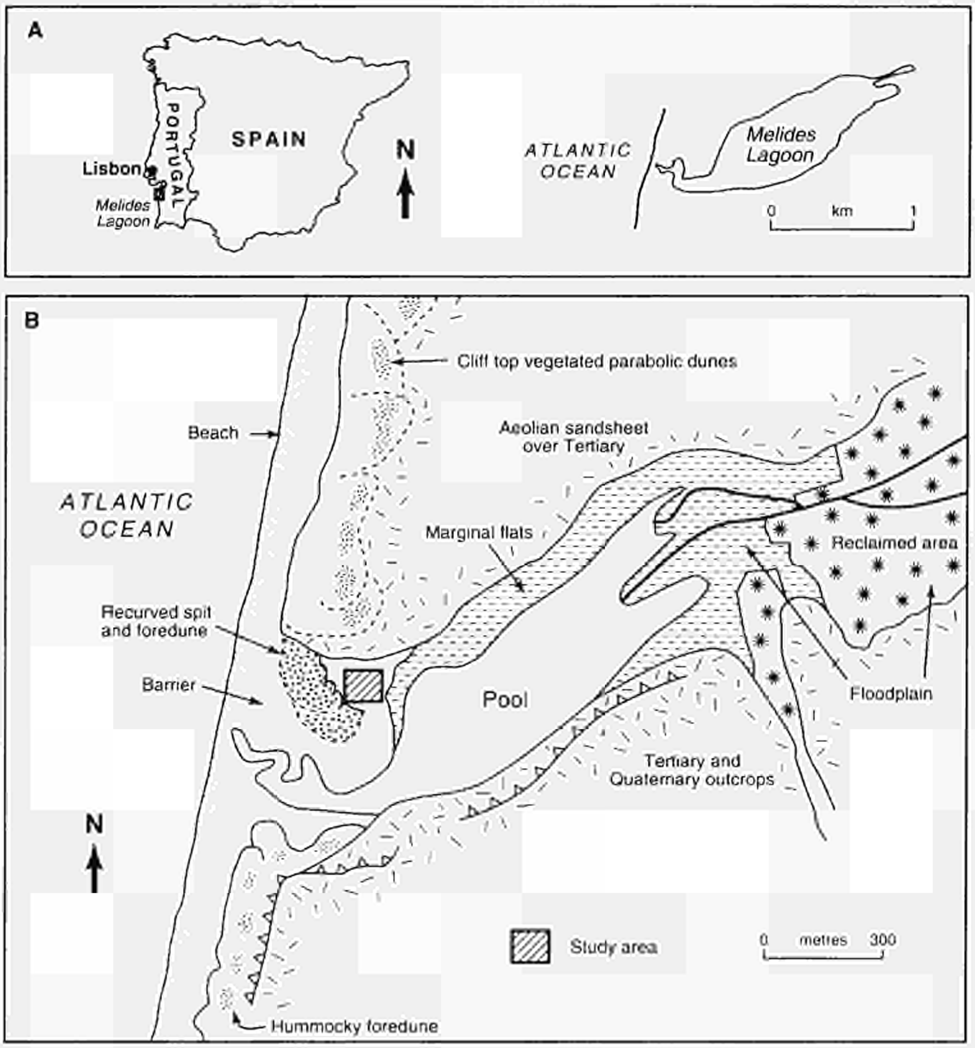


FIGURE 3.7 Melides Lagoon. A - Location map. B - Geomorphological sketch and coring area. C - Location of cores in a cross-barrier transect (Overleaf)

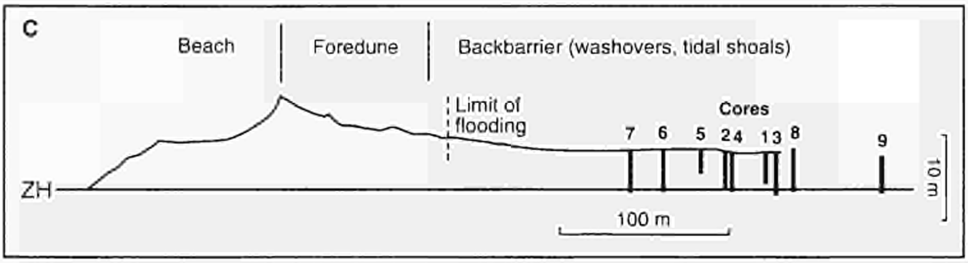


FIGURE 3.7 Part C (see previous page for caption)

Figure 3.8 presents the lithostratigraphic interpretation of the Melides backbarrier structure. Three main lithostatigraphic units were found to represent *circa* 4500 radiocarbon years of sedimentation, arranged in a transgressive sequence:

Lower unit (C) - The lower unit corresponds essentially to alternating layers of sand and muddy sands. The palaeoecological results suggest that the base of this unit was deposited in a low-energy, deep open-water lagoonal environment dominated by fresh or slightly brackish water, at *circa* 4500 BP. This environment is quite similar to the present-day conditions of the inner lagoonal pool during persistent closed-inlet periods. This pattern of sedimentation was replaced in the remnant of the basal unit (4500-3700BP) by coarse-clastic marine inputs related to episodes of breaching of the barrier, that must have been relatively permeable for long periods of time.

Intermediate unit (B) - This sedimentary unit was further divided in sub-units B1 and B2 on the basis of textural contrast:
Subunit B2 - This subunit consists of sandy mud and, to a less extent, of muddy sand, with a high organic content.
Subunit B1 - This sub-unit is completely barren of shells and consists essentially of mud and secondarily of sandy mud. Transition to the upper unit A is quite sharp.
Sand sediments from the intermediate unit had a terrestrial source, either fluvial or derived from overland flow. They show strong textural and geochemical affinities with the Tertiary outcrops that surround the present-day lagoonal basin and are believed to represent the major sediment source for the lagoon. Sedimentation of this unit occurred between 3700 BP and 740 BP in a low-energy environment, subjected to important hydrological fluctuations and included a number of very large, fresh to slightly brackish-water flooding episodes, particularly at the base of sub-unit B1. The palaeoecological record indicates that the lagoonal space evolved as a closed-inlet situation, the silting-up being essentially ensured by terrestrial inputs and in-situ biological production in a progressively more restricted and shallower environment. This pattern was occasionally disturbed by short episodes of marine inundation, related to ephemeral episodes of open-inlet sedimentation or storm overwash of the barrier.

Upper unit (A) - This unit (740BP till present) consists of clean, medium to coarse marine sand with some finer sand at its base and free of any organic matter. It shows strong textural affinities with present-day beach-berm and overwash sands; a foredune facies was also recognised in its basal section. The bulk of unit A was interpreted as being of a marine nature and representing a massive episode of barrier retreat or extension over the outer lagoonal rim.

The interpretation of the stratigraphy of the Melides barrier suggests that the evolution of this coastal sedimentary environment throughout the last 4500 years was characterised by a succession of punctuated equilibria, similar to the model depicted in Figure 3.5D, above. The major lithostratigraphic units represent metastable sedimentation episodes that typically last circa 10^3 years. The long-term trend that characterises each metastable sedimentation episode relates to

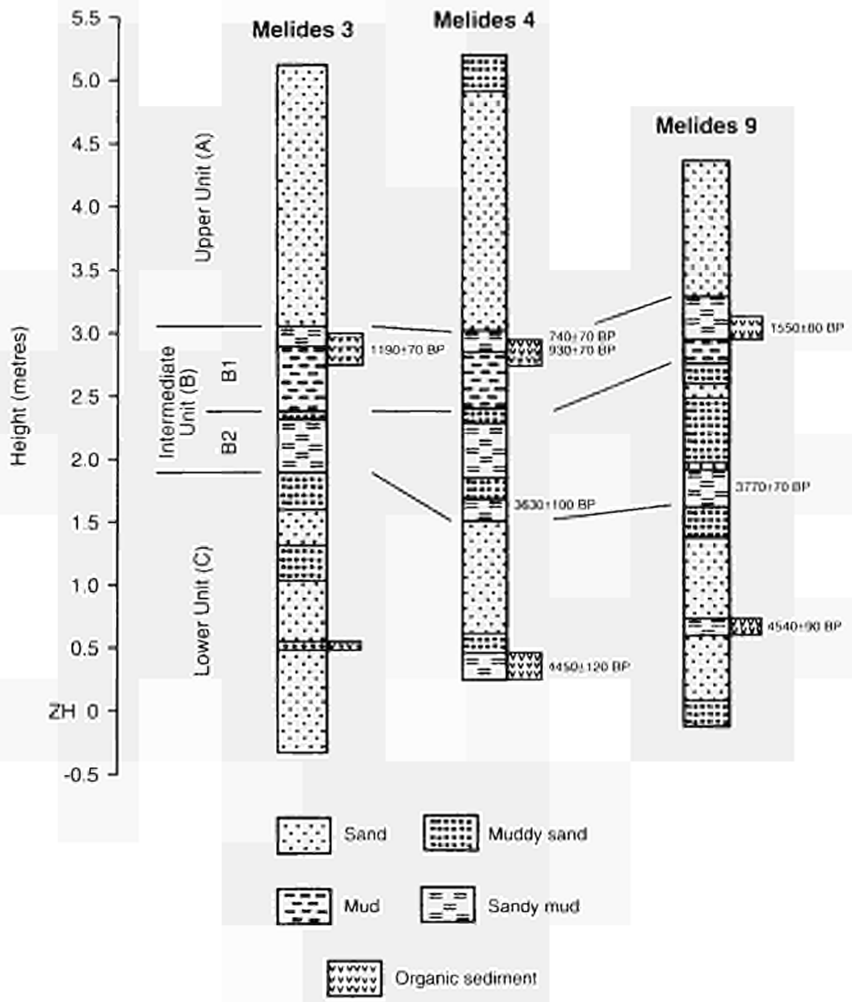


FIGURE 3.8 Lithostratigraphy of the Melides barrier

persistent sea level change and is indicated by the transgressive nature of the whole sedimentary sequence.

Each two contiguous sedimentation episodes are separated by sharp discontinuities. A discontinuity represents the breaching of a threshold and implies rapid change on the geological timescale and the setting of a new equilibrium. This can either revert to a former one (cf. the base of the intermediate unit) or, as illustrated by the setting of the top unit, lead to an irreversible change towards a new level of equilibrium.

In this interpretation rapid changes are as important or even exceed the importance of the established long term-trend of the whole process of coastal evolution through time. They result of the exceedence of local thresholds that are not necessarily associated with single high-energy events. Actually, they may represent no more than small changes of the forcing parameters regulating local coastal equilibria (e.g. changes of the wind or wave regime) or the steady accumulation of small perturbations in the local sediment budget. The result will be accumulation of stress that in a certain moment will exceed the absorption ability of coastal resilience, generating a catastrophic response.

Albufeira. The Albufeira coastal lagoon is located on the western Portuguese coast, some 20km south of Lisbon (Figure 3.9A and B) and lies at the mouth of the Ribeira da Apostiça valley. The Holocene transgression flooded this valley system that silted up extensively leaving the actual lagoon as a relict sedimentary environment. The lagoon is partly closed, according to Nichols & Allen (1981) with a flooded surface of 1.3km^2. It is sheltered from the ocean by a reflective coarse-sandy barrier (Figure 3.9 part D) and is made of two major basinsseparated by a narrow and shallow channel (Figure 3.9 part C). The main body (Lagoa Grande) is partly segmented according to the model of Zenkovitch (1959).

The present-day morphological and sedimentary dynamics of this lagoon and associated barrier was extensively studied by Freitas *et al.* (1992), Freitas & Andrade (1994) and Freitas (1995). This knowledge has been used to interpret the stratigraphical, sedimentological and diatom records of the Late Holocene sedimentary infill of this lagoon using data from a 7.60 m long core collected in the Lagoa Pequena Basin, that accounts for the last 2500 years of sedimentation (cf. Bao *et al.*, in press). The pollen analyses by Queiroz (1989) and Queiroz & Mateus (1994) and the work of Freitas
. (1993) and Freitas (1995) were used to complement this information and to evaluate the impact of human settlements in the area.

The terminal barrier was formed and persisted as an important environmental control since circa 5000 BP and this is probably the only dominant sea level index preserved in this sediment sequence - actually a marked decrease of the sea level rise rate happened at this time, allowing the barrier to develop. During the last 2500 years eustasy was clearly a minor environmental control on the sediment dynamics of this lagoon and associated barrier. This dynamics have been essentially forced by local factors, such as changes in the barrier permeability. These changes may be observed in present-day analogues and control microscale morphological changes, sediment fluxes and water quality and depth across the lagoon.

Within the sediment column (Figure 3.10) three threshold responses may be recognised. They correspond to (rapid) changes in the permeability of the barrier that dramatically changed the volume of water exchanged between the lagoon and the ocean. These changes affected also the nature of sediment supply, the sedimentation rates and both the water quality and depth within the lagoonal area.

The first threshold (circa 2370 BP) did not produce a specific lithological signal but implies an almost permanent isolation of the lagoonal basin and long-lasting freshwater flooding. A second threshold was identified circa 1600 BP when a dramatic opening of the tidal inlet changed the sedimentation regime across the whole lagoonal area, which became dominated by inorganic, minerogenic sediments. Finally, a third threshold has been identified at *circa* 1225 BP corresponding to the establishment of the present-day fluviatile-dominated sedimentation. Both the sedimentological features and the diatom associations indicate the establishment of a more tidal influenced environment due to regular artificial breaching of the barrier, associated with increasing anthropic intervention in the system.

3.2.1.3 High energy coastal flooding and coastal change

Rapid coastal changes triggered by high-energy, low frequency events, such as flooding associated with tsunamis or extreme storm surges, are an accepted factor in the coastal environment. The amount of energy released by such events during a short period may exceed the mesoscale modal energy-levels of any coast by several orders of magnitude. Not surprisingly, rapid and catastrophic responses of the coast can take place, eventually including a number of irreversible changes in the geomorphology of the coast or in its sediment budget.

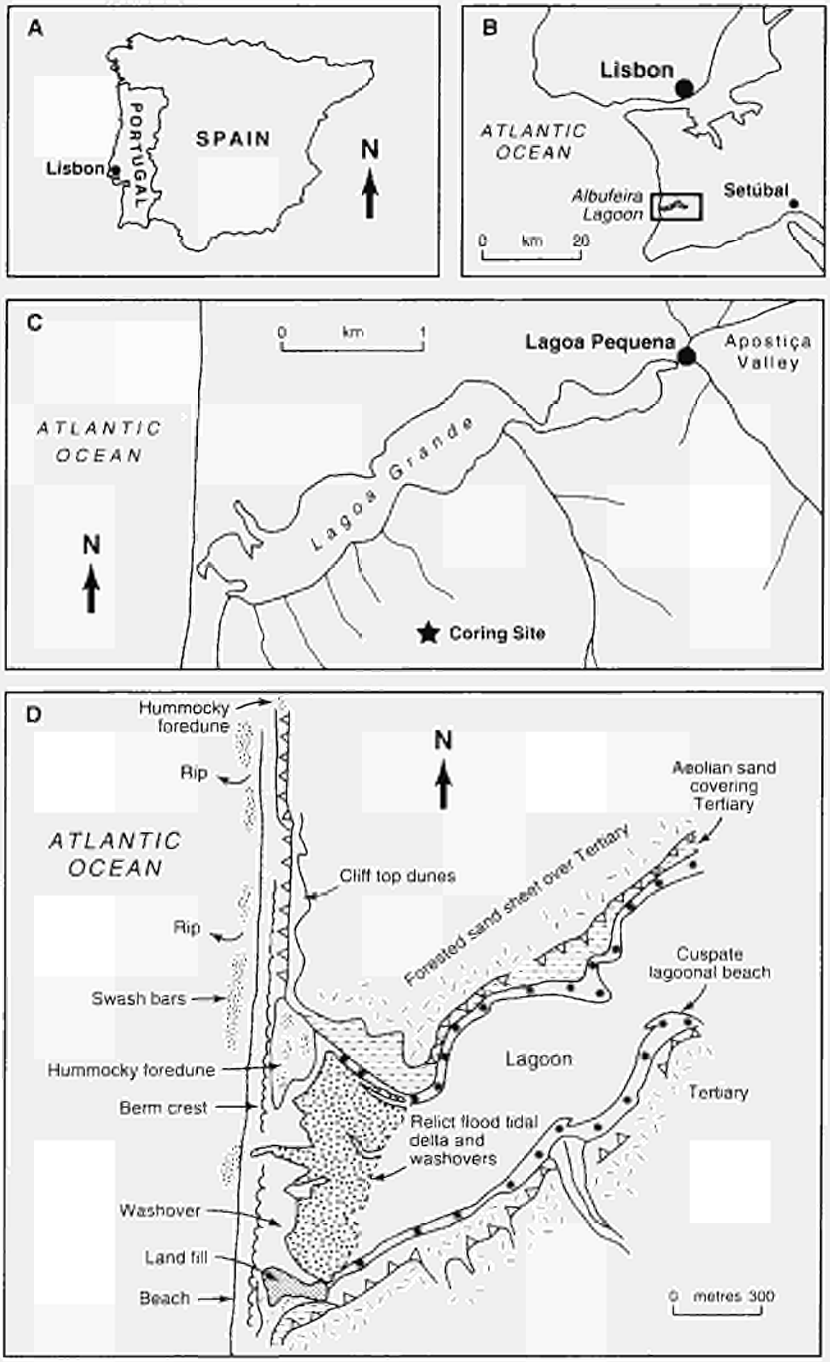


FIGURE 3.9 Albufeira Lagoon. A and B - Location maps. C - Detail of the lagoon and coring site. D - Geomorphological sketch of the barrier and lagoonal environment.

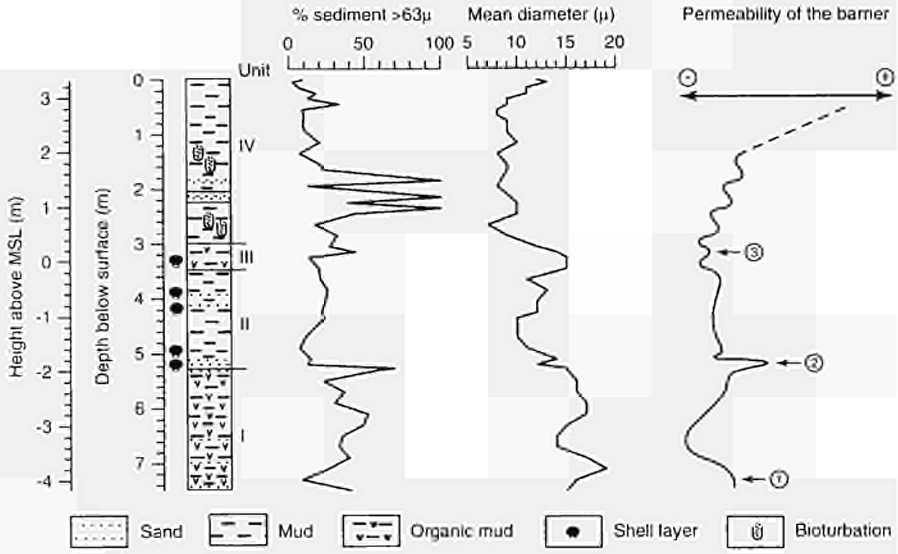


FIGURE 3.10 Sedimentary units and textural parameters of the late Holocene infill of the Apostiça valley (Albufeira lagoon). An interpretation of the changes of the permeability of the barrier throughout the column is shown (1, 2, 3 - thresholds)

Rapid coastal changes may also be triggered by more subtle changes in environmental controls. Examples of these are the changes of sea level rise rate, variation of the sand supply rate to a certain element of the coast or climate-driven changes of the local wind and wave regime. The most important requirement for the occurrence of a rapid change is that the resilience of the coast is not able to accomodate the stress-level generated by the variation of a single or several environmental controls leading to the exceedance of some equilibrium threshold.

Rapid coastal changes related to high-energy single events (or low-energy disturbances of the coastal equilibrium) by natural or man-induced causes are well known in coastal society, well documented in the historical record and an object of concern to coastal managers and engineers. This implies that these changes are neither rare nor specific to any particular coastal setting. Surprisingly, they are scarcely identified and interpreted as such in the geological record, especially in the Holocene coastal sedimentary record. In fact, most of the research undertaken to date in coastal sedimentary sequences is directed at understanding and linking past coastal changes to the activity of long-period driving forces, such as sea level or climate change. This knowledge has been used to build black-box models of coastal response to the expected near-future modification of the global climate setting or sea level rise rate. Contrasting with the amount of work already invested in this type of model, very little is known about coastal resilience, environmental thresholds or rapid coastal processes and, in particular, about their relative contribution to the process of longer-term coastal change.

The case studies presented here illustrate the importance of rapid changes in the modification of the coastal sedimentation regime, coastal geomorphology or morphodynamics in a suite of timescales, ranging from the micro- to macroscale. Some of these changes may be reversible but most of them are not, implying that the exceedence of a certain threshold will result in the rearrangement of new equilibria for the coastal system. Part of these short-term changes may be ascribed to the activity of high-energy, low frequency events of coastal flooding. They are able to damage and reshape the coastal fringe, producing a different physiographic setting that will feed-back with all the components of the coastal system. This is the case of tsunamis and, by extension, of extreme storm surges. Plausibly, the same type of effects may occur if high-energy storms increase in frequency and intensity (as available global climate models forecast for the northern hemisphere and the european littoral fringe) exceeding the reorganisation capacity of the coastal system.

However, the examples quoted here clearly indicate that equilibria thresholds may also be exceeded without the need for catastrophic forcing. Slight changes of the wind and wave regime or anthropogenic intervention in the shore, climatic driven changes of the plant cover or the accumulation of small-scale coastal readjustments may induce disturbances of the sediment budget of any coastal location. The effects associated with these small modifications accumulate until coastal resilience and some particular threshold of equilibrium are exceeded, leading to a rapid coastal readjustment that may well remain recorded in the sedimentary column.

It is believed that a thorough re-examination of the Holocene sedimentary sequences around European shores and their re-interpretation, based on the careful examination of present-day analogues is urgently needed. This research would certainly highlight the importance of rapid coastal changes to the overall modification of the coastal system and introduce an innovative perspective in models of future coastal change. Finally it would help to close the gap between the microscale approach, typical of engineering and management studies and the meso- to macroscale approach, common to earth scientists.

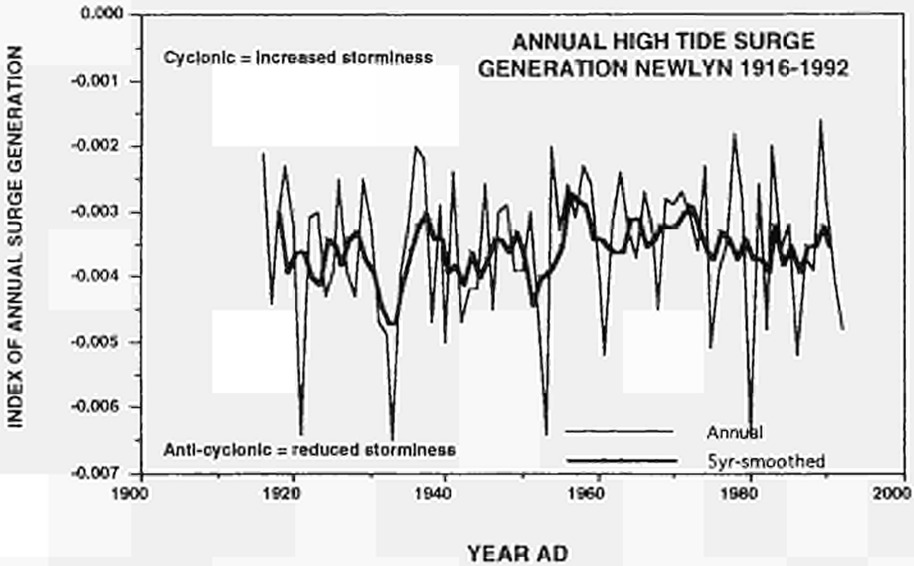
3.3 The spatial and temporal patterns of climate-related coastal flooding

Alastair G. Dawson & David E. Smith

Recent years have seen increased interest in investigating patterns of extreme flooding on various timescales in an effort to determine relationships with weather patterns and identify trends which may relate to pertubations in the climate system. Two broad approaches have been used. For timescales of up to 500 years, studies of documentary and instrumental evidence have been employed, whilst for mainly longer timescales geological evidence has been examined. This section reviews some recent studies illustrative of the approaches involved.

3.3.1. Documentary and instrumental evidence

A number of studies have attempted to determine trends in coastal flooding on a century scale from instrumental evidence. For example, Orford *et al.* (1996b) examined storm surge activity from hourly tide gauge values at an annual level of summary for Newlyn (UK) and Brest (France) (Figure 3.11). Data were detrended for secular sea level change and characterised according to annual surge value and frequency. The Index of surge activity is the slope coefficient of a negative regression linking log surge frequency with surge height. The lower the absolute value of B1 the higher the likelihood of storminess leading to enhanced surge generation. The smoothing of these data using a 5-year moving avereage highlights a sub-decadal oscillation in storminess (5-7 years) superimposed on longer-term cycles that last over several decades. These scales of oscillation are even more apparent in the data structure for Brest which identifies a sequence of major cycles of storminess variation (decadal scale) in the later part of the 19^{th}-century. The level of variation of that time exceeds the scale of variation identified in the later part of the 20^{th}-century. It is interesting that these data while suggesting that surge-generating storminess in the late 20^{th}-century is consistently greater than in earlier parts of the century (see also Bouligand and Pirazzoli, 1999), does not warrant the statement that annual storminess variation is greater now than in the past. The coherence between the Brest and Newlyn signals given their spatial closeness should not be surprising, but some of the differences between the records may reflect spatial differentials of each gauging station to extreme event depression tracks. These oscillations in storminess are a first stage in the probabilistic process of identifying the scale of key diagnostic variables that coincide to set the conditions by which extreme storms occur.



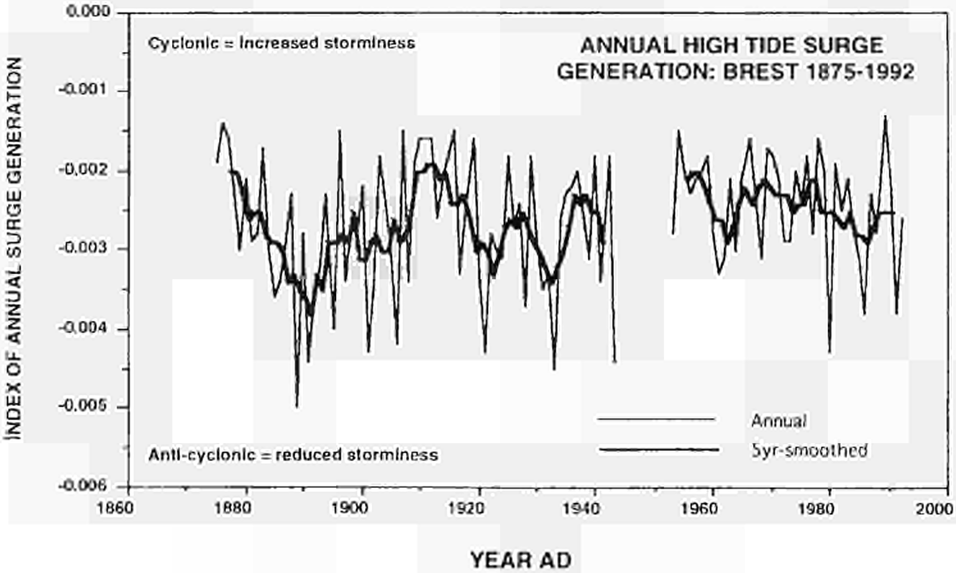


FIGURE 3.11 De-trending of hourly tide gauge values for Newlyn, UK and Brest, France.

Century-scale studies may not, however, identify longer period fluctuations in storminess and coastal flooding. Consequently many studies have adopted a longer view, accepting that the data may be less secure. Much of this work has built upon or been influenced by the detailed studies of writers such as Gottschalk (1971, 1975, 1977) and particularly Lamb (e.g. 1977, 1984a, 1984b, 1991), who recognised that in many European countries a rich archive of anecdotal and early instrumental information on weather is available, complemented across Europe by a large number of stations where modern weather records have been made. Some examples of this work are as follows:-

Camuffo (e.g. 1993) has analysed records of storm surges at Venice from AD782 to 1990, the period up to 1800 based upon documentary evidence and the period since based first upon early instrumental evidence then later (since 1923) upon modern instrumental evidence. Camuffo found that particularly high surge levels have been taking place since 1914 and discussed the factors involved, concluding that anthropogenic factors have been strongly influential. In Ireland, Tyrrell and Hickey (1991) combined documentary and instrumental evidence to present a flood chronology for Cork for 1841 to 1988, demonstrating an increase in flood events during the present century, whilst in Scotland, Hickey (1998) presented a detailed chronology of coastal flooding from 1500 to 1991 (Figure 3.12). He identified periods of increased coastal flooding between 1620 and 1700 (the peak of the "Little Ice Age") and from 1850 to 1875. Also in Scotland Dawson *et al.* (1997) presented an analysis of a 200 year record of gale frequency for Edinburgh from 1780 to 1988, probably the longest historical record of gales in Europe (Figure 3.13). They determined three clear peaks in storminess following the volcanic eruptions of Tambora (1815), Krakatoa (1883) and El Chichon (1982) and concluded that the greatest periods of storminess in the Edinburgh record followed episodes of volcanism.

Hulme & Jenkins (1998) show the changing pattern of gale activity purporting to represent the United Kingdom for the historical period. They note that former gale activity has been highly variable with a distinctive minimum occurring in 1985 and a maximum in 1887. They argue that for Britain, the 1961-1990 average is for just over 12 severe gales per year, mostly between November and March, and they note also that the middle decades of the twentieth century were rather less affected by severe gales than the earlier and later decades. Hulme & Jenkins (1998) concluded that he most recent decade (1988-1997) was characterised by the highest frequency of severe gales (15.4/year).

In a study of gale day frequency for Edinburgh for the period 1767 -1993, Dawson *et al.* (1997) showed that historical patterns of gale frequency have exhibited complex variability. The most notable features of the analysis were:

a) that the periods between 1811-1819 and 1872-1890 were by far the stormiest episodes that have occurred in recent times.
b) that the recent decade of stormy weather between 1988-1997, although the most severe period of storminess during the 20th century, was considerably less severe than during 1811-1819 and 1872-1887. For example, the gale day frequency for Edinburgh for 1815 (57), 1816 ((67), 1817 (56) and 1818 (72) can be compared with the two stormiest years of recent years 1982 (47) and 1983 (42).
c) that there have been periods when storms, as represented by gale days, have been relatively infrequent, for example the decade 1910-1920.

The most important result of the latter analysis is that the pattern of gale day frequency for Edinburgh bears little relationship to the results of Hulme & Jenkins (1998). Whereas these authors present average vales for the UK as a whole, the Edinburgh data shows strong departure from these UK average values. More recent (unpublished) analysis of historic storm data for the Shetland Isles and Stornoway, Outer Hebrides show that the Hulme & Jenkins (1998) pattern of UK-average storminess bears little relation to empirical data on historic storm frequency for Scotland. Illustrative of this are the gale day frequency values for Stornoway which during 1888 reached as high as 115 (compare with the Hulme & Jenkins average for 1988-97 of 15.4 gale days/year)!

North Atlantic Oscillation (NAO)

Hurrell (1995) and Appenzeller *et al.* (1998) devised an index of air pressure difference between the Azores high and the Icelandic low and attempted to quantify this pattern of air circulation in respect of a North Atlantic Oscillation (NAO) Index that they considered as a reliable indicator of climate variability in the North Atlantic region. Periods of strongly positive NAO corresponded with a strong Azores high and a deep Icelandic low. Conversely, periods when the NAO Index was negative corresponded with the times when the Icelandic low was poorly developed (or not developed at all) and the Azores high was replaced by lower than average air pressure. A strongly positive NAO Index, therefore, corresponds with periods of time coincident with increased storminess around the Scottish coastline. These changes in the NAO Index have been reconstructed from *circa* 1650 AD till present and they show a high degree of temporal variability (Appenzeller *et al.*, 1998). Comparison of the NAO Index values since 1824 with records of gale day frequency from Edinburgh (Dawson *et al.*, 1997) show good agreement. Thus, periods of increased storminess generally correspond to periods of time when the NAO was strongly positive.

The processes that link the NAO to storminess in (e.g) Scotland are extremely complex, however. For example, there is reason to believe that changes in the NAO are also linked to changes in North Atlantic thermohaline circulation. Thermohaline circulation in the North Atlantic depends on the continued sinking of surface waters in the Greenland Sea. This process is accomplished through the influence of cold katabatic winds from the Greenland ice sheet that caused the ocean surface to freeze and produce sea ice. Sea ice formation is accompanied by the expulsion of salt that causes the adjacent waters to sink and thereby produce Deep Water.

During periods when the NAO is strongly positive, easterly winds blow across the ocean surface towards Greenland and disrupt the process of Deep Water formation thereby slowing down the rate of thermohaline circulation in the North Atlantic. Since thermohaline circulation in the North Atlantic is the principal process by which heat is released to the atmosphere (more commonly described as the Gulf Stream), it can be argued this process is by far the most important mechanism that influence climate change in NW Europe.

The global modelling of climate change as described for example in the IPCC96 (Warrick *et al.*, 1996) and in the UKCIP report (Hulme & Jenkins 1998) does not take account of such processes in the simulations of future climate change. This is a major failing of the models, these being generally restricted to simple 1-D upwelling-diffusion model calculations. Since the present numerical models of climate change are not yet sufficiently advanced to incorporate the complexities of North Atlantic thermohaline changes on the climate of NW Europe, they cannot hope to provide accurate predictions of how the global atmospheric warming will affect the climate of NW Europe.

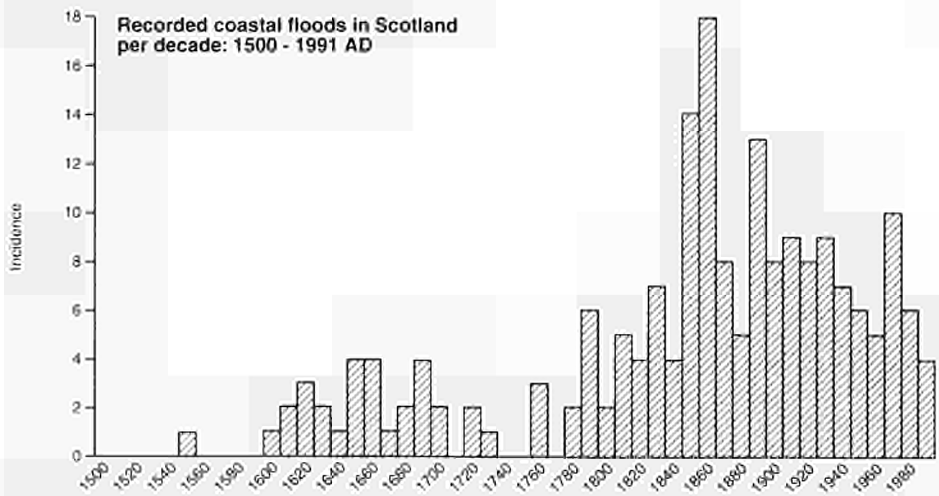


FIGURE 3.12 Recorded coastal floods in Scotland per decade, 1500-1991 (after Hickey, 1998).

Historical documentary records for NW Europe provide historical information on episodes of marine flooding and individual storm surges and wind storms that have taken place during the last *circa* 2000 years (Britton, 1937). The historical observations of such phenomena demonstrate a marked variability in climate in northwest Europe during the last 2000 years. For the time period prior to documentary records reliance has to be placed geological investigations. Such studies (e.g. Goodbred and Hine, 1995) have highlighted the particular difficulties in deciphering former storm surges within the geological sedimentary record. Other studies have highlighted the complex relationship between cyclone genesis and palaeooceanographic changes in the North Atlantic that define the latitudinal tracks of North Atlantic cyclones and the positions where they reach the coastline of northwest Europe. Such changes appear to be closely related to the North Atlantic Oscillation (Hurrell, 1995) that in turn appears to be related to the rate of production of North Atlantic Deep Water and the consequent position of the polar oceanographic front.

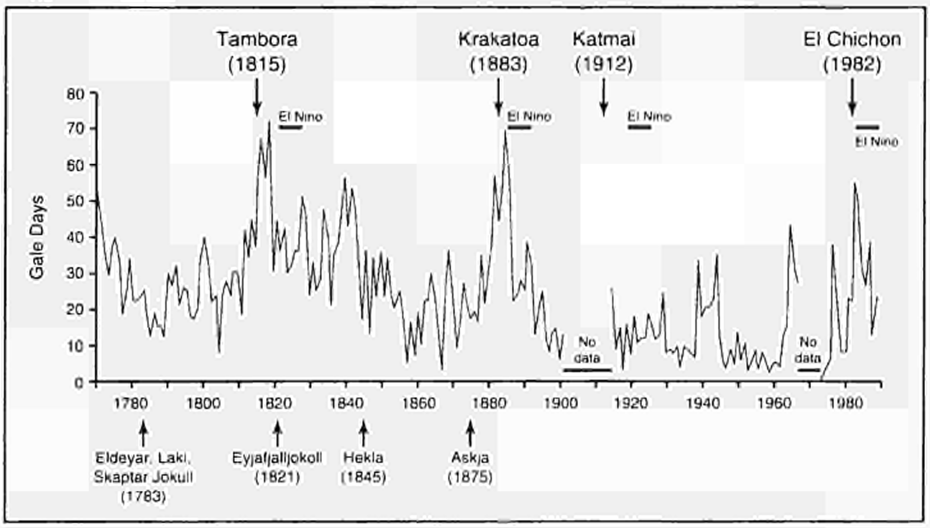


FIGURE 3.13 Gale day frequencies for Edinburgh, UK, 1780-1988 (after Dawson et al., 1997).

Some examples of the more destructive storm surges to have affected NW Europe include in particular two notably destructive storms which took place in January 15th 1363 and in November 26th 1703. Numerous letters testify to the 1363 gale and these are summarised by Britton (1937). Most of the accounts describe the effect of the storm in southern Britain and parts of Ireland. By contrast the storm of 7-8th December 1703 was characterised by a 300 nautical miles wide belt of exceptional destruction across southern England and Wales, the southern North Sea, Netherlands, north Germany, Denmark and possibly also parts of France and Sweden. Many have suggested this is the severest storm of which we have an accurate account. The destructive effects are described in Lamb (1991). The storm appears to have represented the combination of a long-continued cyclone regime that was located adjacent to a winter anticyclone over Scandinavian and possible high pressure over France. This resulted in the concentration of cyclonic activity of the British Isles in the first 6 days of December. Lamb (1991) estimates that the strongest gradient winds may have reached 150 knots with the stronger surface winds in the order of 90 knots or slightly more. Consideration has also been given that the storm represented the remnant of a Caribbean hurricane. The 1703 storm was not reported as exceptionally destructive in northern Scotland. In fact some of the most damaging storms in living memory in northern Scotland (e.g. 1697, 1900) are not reported from weather accounts from southern Britain. These differences highlight the difficulty in identifying trends and patterns in the tracks of individual north Atlantic storms.

Recent research has attempted to explain these differences in terms of the North Atlantic Oscillation (NOA) (Hurrell, 1995; Wilby *et al.*, 1996). This index was developed on the basis of analysis of historic records of air pressure variations at numerous sites throughout the Atlantic seaboard of Europe and conventionally measured as the difference in air pressure between two stations in Iceland and Portugal respectively. During periods of exceptionally low pressure over Iceland and at times when the Azores high is particularly strong, the difference in air pressure between the two ridges is at its highest and has been described by Hurrell as equivalent to a strongly positive North Atlantic Oscillation index value of +1. By contrast, time periods associated with exceptionally high air pressure over Iceland and an exceptionally weakened Azores high, correspond with an extremely negative North Atlantic Oscillation index of -1.

The temporal changes in the NOA index since 1859 were explained by Hurrell as responding to changes in the position of the polar oceanic and atmospheric fronts in the North Atlantic and the location of this boundary zone was of the greatest importance of defining the latitudinal position of the circumpolar vortex and hence the zone within, southward of which North Atlantic cyclones were likely to be generated. Hurrell recognised that meridional changes in the tracks of North Atlantic cyclones had taken place during the period of meteorological records (since 1859) and his interpretations were investigated in detail in respect of the UK by Wilby *et al.* (1996). The latter authors demonstrated that north-south changes in the location of North Atlantic cyclones could provide a reasonable account for regional variations in historic climate and could explain the occurrence of periods of storminess in some areas yet not in others.

The precise causes of changes in the North Atlantic Oscillation index are unclear, however, and may include or be influenced by changing rates in the production of North Atlantic Deep Water, changing intensity of the North Atlantic Gulf Stream in turn influenced by deep water formation as well as other effects including changing Rossby wave lengths, effects of El Niño and influences caused by a major volcanic eruptions together with solar changes.

3.3.2. Geological evidence

David E. Smith

In seeking to determine the relationships between flooding events and known climate variations, as well as to extend the documentary and instrumental record back in time to further identify trends, many recent studies have turned to geological evidence. This approach is unlikely ever to match the detail with which storminess and coastal flooding can be determined from instrumental and sound documentary evidence, but improvements in techniques in recent years have enabled a number of useful contributions to be made. Such improvements include much more detailed field work, both morphological and stratigraphical, including the recognition that a broad 'sampling' approach is not always appropriate, recognising variations in the coastal environment. Relating evidence to a common datum is also now more widespread, with an acceptance that features and deposits need to be related to particular states of the tidal cycle (the 'indicative meaning' approach of Tooley (1982)). Microfossil studies are now undertaken in more detail than previously and studies commonly involve two or more techniques, whilst a greater range of absolute dating methods are used. In reconstructing flooding events from the geological past, two broad categories of events are identified in many European studies, the secular change where flooding over a wide area of greater than regional dimensions is identified, and regional or sub-regional changes, where more localised events are recognised.

3.3.2.1. Secular changes

It is generally recognised that deglaciation of the last major ice sheets was followed by rapid, meltwater-induced rises in sea surface levels (e.g. Fairbanks, 1989; Blanchon and Shaw, 1995).

Whilst the detail remains unclear, a strong rise in sea surface levels, possibly in response to meltwater discharge, took place during the early and middle Holocene, causing widespread and rapid flooding between *circa* 8500 ^{14}C years BP and *circa* 7000 ^{14}C years BP along many European coastlines (see for example Pirazzoli, 1996 & 1991). The amount of sea surface rise outside areas of strong glacio-isostatic uplift at that time amounted to up to at least 25m. In the Limfjord region of Denmark, Petersen and Rasmussen (1995) identified a rise of 28m starting at *circa* 8600 calendar years BP and achieving a rate of 33mm a^{-1}, whilst in the UK, Zong and Tooley (1996) estimated a rise of *circa* 11m between 8870 and 8510 calendar years BP at a maximum rate of 36.7mm a^{-1} for Morecambe Bay, NW England, UK whilst Smith *et al.* (1999a) estimate a rise of *circa* 10.5m between 9478 and 7871 calendar years BP at a maximum rate of 9.97mm a^{-1} and a mean rate of 7.09mm a^{-1} for the Ythan valley, NE Scotland, UK (Figure 3.14).

In Belgium, Baeteman *et al.* (1999) estimate that in a period up to *circa* 7500 calendar years BP, relative sea level rose by an average rate of 7mm a^{-1}, and similar rates have been estimated for this period elsewhere around European coastlines (see for example Pirazzoli, 1991). These and other estimates probably reflect the third meltwater discharge of Blanchon and Shaw (1995), and although more detail is awaited it seems clear that the middle Holocene saw rates of sea surface rise around European coasts which resulted in rapid coastal change, and that only in areas rising isostatically (notably much of Scandinavia and central Scotland) was the rise either absent or sufficiently small in rate and amount that coastal geomorphic processes could temper the widespread inundations taking place elsewhere.

In one respect the early to middle Holocene rise was unique, and is unlikely to be matched by future sea surface rises (unless rapid disintegration of the West Antarctic ice sheet takes place), but towards the margins of glacio-isostatically rising areas in Europe, the modified rises are analagous to forecasts of future rises due to global climate change.

3.3.2.2. Storm surges

In recent years, a number of studies have identified storm surge deposits in the geological record from a number of coastal settings. Examples of these are as follows:

Rocky coasts. From NW Scotland, UK, Shennan *et al.* (pers.comm.) believe that storm surge signatures may be obtained from isolation basins, although no consistent record has so far been obtained. Further north, in Orkney and Shetland, the cliff-top accumulations recognised by Hall (e.g. 1996) and referred to earlier are good evidence of storm surge activity.

Dune-fringed coasts. These coasts, usually exposed, have a considerable potential to record storm surge activity. Episodes of dune development, such as have been recognised by Jelgersma *et al.* (1970) for the Netherlands and Tooley (1990) for the United Kingdom, may reflect increases in storm surge activity, although other explanations are possible, as have been pointed out in Mac Clenahan's (1997) excellent review. More locally, Jelgersma *et al.* (1995) claim to identify storm surges from layers of shells in coastal dunes, whilst Keatinge and Dickson (1979) interpret the extension of blown sand deposits in Bay of Skaill, Orkney, UK at *circa* 5000 ^{14}C years BP as reflecting increased storminess. Gilbertson *et al.* (1999) identify episodes of sand drift in peat mosses within and landward of coastal dunes in the Outer Hebrides, Scotland, UK, whilst Smith *et al.* (1999b) have dated a prominent episode of dune expansion in the same area to the start of the storminess associated with the Little Ice Age. In NW Spain and western Ireland, Devoy *et al.* (1993) report episodes of flooding recorded in peat and estuarine environments landward of dune systems, as does Smith (1993) in the Shetland Islands, UK.

Back barrier environments. Sand and gravel levels in lacustrine or peat deposits landward of barriers have been interpreted as evidence of storminess and flooding episodes. At Lochhouses, eastern Scotland, UK, Smith *et al.* (1991) identified events at 6930±75, 6760±75, 6370±70 and 6035±70 ^{14}C years BP, whilst in Brittany, France, Regnauld *et al.* (1996) describe washover lobes

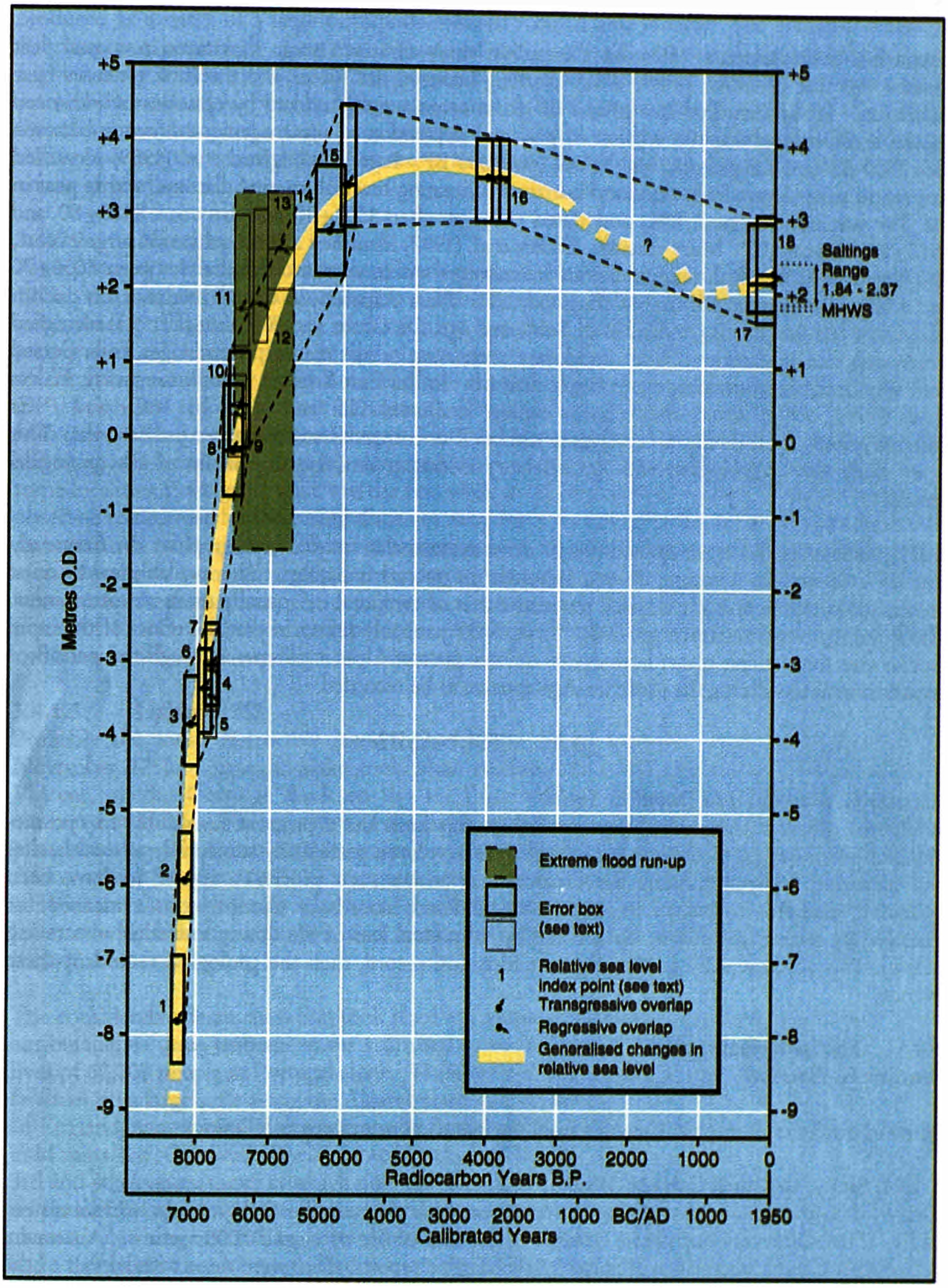


FIGURE 3.14 Relative sea level rise during the the Main Postglacial transgression in Northeast Scotland, UK. Reproduced by permission of the Royal Society of Edinburgh, (Smith et al, 1999a).

and sand horizons behind barriers, reflecting flooding events occurring on several occasions during the late Holocene.

Estuarine environments. Possibly the earliest study of storm surge signatures in coastal peat mosses was that of Linke (1979, 1981), working between the Weser and the Elbe estuaries near Cuxhaven. Linke identified two phases of storm surge activity marked by episodes of enhanced coastal erosion involving the erosion of peat and deposition of minerogenic sediments, between *circa* 2500 BP and the present, and before *circa* 4000 BP. Later, Cullingford *et al.* (1989) identified two storm surge layers from thin sand horizons containing brackish-marine diatoms, within peat in the Tay estuary, Scotland, UK, and dated them at after 8565±80 and between 8485±80 and 8510±85 ^{14}C years BP. More recently, Tooley *et al.* (1997) identified a series of storm surge events, also based upon brackish-marine diatom assemblages within peat and dated at between 6030±70 and 5740±60 ^{14}C years BP in NW England, UK. Near Alkmaar, in the Netherlands, in diatom analysis of the Bergen clay, a deposit of Subboreal age, De Groot and Westerhoff (1993) identified storm deposits consisting of thin sand layers containing robust diatom assemblages, interspersed with silty horizons containing more fragile diatoms. In the East Friesian Islands, Germany, Ehlers *et al.* (1993, 1994) dated shelly layers within laminated silts and clays by ^{134}Cs and ^{137}Cs determinations, establishing a correlation with well documented storm surges in the area. This latter study ties in geological with documentary evidence, verifying the nature of the geological record.

Such evidence as has been quoted above demonstrates that coastal stratigraphies do frequently contain evidence for flooding events, although the record is patchy. The available evidence at present does not permit the detailed reconstruction of temporal or spatial patterns of storminess. The flooding events recorded pre-6000 ^{14}C years BP probably reflect in part the effect of the rapid secular rise then taking place; later events do not present a clear pattern, although the period of greater storminess during the Little Ice Age appears to be recorded.

3.3 TSUNAMIS

Alastair G. Dawson and David E. Smith

Whilst the Atlantic seaboard of western Europe has been the area most susceptible to cyclone-induced storm surges, Mediterranean coastal areas have been particularly susceptible to earthquakes and tsunamis. Notwithstanding the timescales at which such processes appear to have been operative, tsunamis constitute important coastal flood hazards in Europe as is evidenced for example, by recent tsunamis in Greece and by substantial loss of life during individual events (e.g below). This account will consider first the historical record, then the geological record of these events.

3.4.1 The historical record

Alastair G. Dawson

Some examples of European tsunamis from the historical record are as follows:

3.4.1.1. Sicily, 1693 and Calabia, 1783

One of the most severe earthquakes and tsunamis took place on the 11th January 1693 in eastern Sicily. This disastrous earthquake resulted in the loss of life of circa 70,000 victims. A tsunami occurred at Catania and also at Augusta. According to reports, there were three withdrawals of the sea and three major waves. Similarly severe earthquakes and tsunamis took place in Calabria on 5th and 6th of February 1783. At this time, a tremendous earthquake occurred in this area associated with five very strong quakes. Considerable stretches of the coastline of Calabria were badly affected by a tsunami and the sea was reported to have receded and then inundated the shore with recessions and inundations repeated at least three times at intervals of about 10 minutes. At

Messina harbour, quays and buildings were flooded and in one area the sea withdrew for more than 8 yards leaving the sea bottom dry and a lot of fish on the beach. Then suddenly the water came back surpassing the limit previously reached and flooding the coast. Local tsunamis were also produced as a result of a large earthquake-induced rockfall into the sea. The tsunami of the 6th February was particularly disastrous because of the very high number of victims where many, frightened by the earthquake shocks, escaped to the beach and were drowned by waves which reached the roofs of harbour buildings. In excess of 1,500 people were drowned and the tsunami flood level was estimated to have been between 6 and 9 metres. Tinti and Maramai (1996) observed that in one area the tsunami was observed to have been associated with deposition of '...some sand on the ground'.

3.4.1.2 Messina, 1908

A very strong earthquake and tsunami took place in the Messina Straits on the 28th December 1908. The towns of Messina and Reggio di Calabria were completely destroyed together with many neighbouring villages. The area of destruction was about 6000 kilometres^2 and more than 60,000 people died. The earthquake produced a violent tsunami in the Straits of Messina that caused severe damage and many victims. In all places the first observed movement was the withdrawal of the sea (in some places by about 200 metres). Thereafter coastal flooding took place in association with at least 3 large waves. According to Tinti and Maramai (1996) the tsunami lasted many hours and reached its maximum intensity along parts of the Calabrian coast and on the coast of Sicily. In some localities, the biggest wave was the first while at others it was the second. Tsunami runup was observed to decrease for increasing distances away from the epicentre but in the Messina Straits this was obscured by the effects of waves resonance. At Messina the wave height reached 3 metres, numerous boats were damaged, harbour keys were destroyed, walls collapsed and several boats were transported on shore. In certain areas the maximum level reached by the tsunami was in excess of 10 metres above contemporary sea levels, resulting in the destruction of many buildings and considerable drowning (Tinti and Maramai, 1996).

3.4.1.3. Lisbon, 1755

Probably the most destructive tsunami in Europe during historical times took place on 1st November 1755. An earthquake took place offshore circa 200 kilometres WSW of Cape St Vincent, on the Gorringe Bank on the sea floor west of Portugal and attained a magnitude estimated at 8.5Ms. The epicentre of the earthquake was in an area along the Azores - Gibraltar plate boundary that forms the western part of the lithosphere boundary between the Eurasian and African plates (Moreira, 1985). The eastern section of the Azores - Gibraltar plate boundary (which includes the Gorringe Bank) is an zone of active plate compression and in this area faults tend to have a large source component that results in high-magnitude and deep-seated tsunamigenic earthquakes).

The considerable destruction that took place in Lisbon, in addition to widespread fires, was mostly attributable to three tsunami waves estimated to be between 5 and 13 metres high that took the lives of 60,000 people in Portugal alone. There are also numerous reports of tsunami flooding and fatalities on a large scale along the Algarve coast and on the coastline of Morocco (Andrade, 1992). In England, contemporary observations by Borlase (1755, 1758) describe the arrival of the tsunami in Mounts Bay, Cornwall. Borlase noted '... the first and second refluxes were not so violent as the 3rd and 4th (*tsunami waves*) at which time the sea was as rapid as that of a mill-stream descending to an undershot wheel and the rebounds of the sea continued in their full-fury for fully 2 hours... alternatively rising and falling, each retreat and advance nearly of the space of 10 minutes until 5 and a half hours after it began'. Reconstructed tidal changes for this day for the Isles of Scilly show that the time of high tide coincided approximately with the arrival of the first tsunami wave (Foster *et al.*, 1991) some five hours after the first shocks were reported on the Portuguese coast. There are no known reports of the progress of the tsunami northeast along the Channel but it is reasoned here that the coastal flooding effects must have been considerable. There is some evidence to indicate that the 1755 Lisbon tsunami was not solely caused by a sea bed fault. Recently a large turbidite/submarine slide complex has been identified on the seafloor adjacent to the Gorringe Bank and tentatively dated to 1755 (Weaver and Masson, pers. comm). This discovery raises the possibility that the tsunami was partly generated by an earthquake-triggered fault on the sea bed and partly by submarine sediment slumping.

3.4.1.3 Greece

In some cases major earthquakes appear to have taken place and to have caused significant changes to individual coastlines. On occasions such earthquakes have been reported to have generated tsunamis. However, in some cases the evidence for tsunamis is equivocal. A good illustration of that is a period of repeated preseismic and coseismic movements which took place during historical times and reported throughout the eastern Mediterranean as far as the Levant and central Greece. According to Pirazzoli *et al.* (1996) and (1992) many coastal regions were affected by crustal uplift represented by a series of coseismic uplift events which took place during a relatively short period of time between the mid 4th and 6th centuries AD. The regions affected by uplift at this time included several of the Ionian Isles, the eastern Gulf of Corinth, Antikythira Island and most of western Crete as well as parts of southern Turkey, Cyprus, Levant, Syria and Lebanon. Pirazzoli reports that the amount of coseismic uplift was on average between 0.5 and 1.0 metres with at least 9 metres in western Crete (Pirazolli *et al.*, 1996a) where evidence of crustal uplift is particularly clear. Pirrazoli (1996) attributed the clustering of coseismic uplift events to a great subduction earthquake which uplifted and tilted a block of lithosphere approximately 200 kilometre long probably during AD 365 and proposed that this event triggered a series of seismic events elsewhere across a zone at least 1,500 kilometres wide. Such high-magnitude vertical crustal movements imply that such changes are likely to have been associated with a destructive tsunami. There is indeed some evidence of such a tsunami having taken place and having caused destruction as far a field as the Levant. However, the evidence that a large tsunami took place in western Crete (where the uplift was greatest) has been disputed by Dominey-Howes *et al.* (1998).

Historical accounts of former tsunamis have particular value since they can provide information on the frequency and magnitude of events for the time period for which historical records are available. For example in Italy the oldest historical record of a tsunami having taken place is for the 79AD eruption of Vesuvius. Since numerous tsunamis have taken place along the coastline of Italy since then, historical accounts are a particularly valuable archive that can be used to estimate tsunami (and earthquake) recurrence. In recent years however geological investigations have been used to identify former tsunamis that took place in prehistory. This has proved possible owing to the recognition that, in many cases, tsunamis deposit sediment in the coastal zone (Dawson, 1994; 1996). Identification of such sediment layers in coastal sediment sequences has led to a different perspective on the past frequency and magnitude of tsunami events in different European coastal regions.

3.4.2 The geological record

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Geological investigations of former tsunamis is a relatively new research area. The recognition that many tsunamis deposit sediment in the coastal zone has only become an accepted idea during the last 5-10 years. Discussion of this concept has been accompanied by a proliferation of academic papers that have described a range of sediments that have been attributed to a series of former tsunamis (Minoura and Nakaya, 1991; Paskoff, 1991; Atwater, 1992 & 1987; Minoura *et al.*, 1994; Nishimura, 1994; Pinegina *et al.*, 1997; Ranguelov, 1998).

Unlike storm surges, tsunami runup across the coastal zone is frequently associated with the rapid lateral translation of water and suspended sediment. Thus tsunami deposits can be used to provide an indirect record of former offshore earthquakes and underwater landslides. It is exceptionally difficult, however, if not impossible, to differentiate tsunami deposits attributable to former submarine slides or offshore earthquakes (e.g. Perissoratis and Papadopoulos, 1998). In particular areas of the world, especially in areas of an active plate motion where an offshore earthquake has taken place, it may be a gross over-simplification to attribute the triggering mechanism solely to earthquake-induced sea bed faulting. Frequently, an offshore earthquake may also generate local submarine slides thus leading to complex patterns of tsunami flooding at the coast (e.g. Yeh *et al.*, 1993). In other areas (e.g. Hawaiian islands, Norwegian Sea) submarine sediment slides may be the dominant mechanism of tsunami generation (Moore and Moore, 1988; Bondevik *et al.*, 1997).

Tsunami deposits are distinctive (Dawson *et al.*, 1996; Bourgeois and Minoura, 1997). They are frequently associated with the deposition of continuous and discontinuous sediments sheets across large areas of the coastal zone (Dawson *et al.*, 1995). Frequently they consist of deposits of sand containing isolated boulders. On occasions such boulders exhibit evidence of having been transported inland from the nearshore zone. In addition microfossil assemblages of diatoms and foraminifera contained within sand sheets may provide information of onshore transport of sediment from deeper water (Hemphill-Haley, 1995a, 1995b, 1996; Dominey-Howes, 1996a & 1996b).

Field observations of tsunami flooding usually describe the rapid lateral translation of water across the coastal zone. Frequently the lateral water motion associated with runup is influenced by local wave resonance. Thus the tsunami waves as they strike the coast are unlike waves associated with storm surges since not only are they associated with considerably greater wave lengths and wave periods, but they are essentially constructive as they move inland across the coastal zone (Reinhart and Bourgeois, 1989). The rapid water velocities (provided that there is an adequate supply of sediment in nearshore zone), are in most cases associated with the transport of a variety of grain size ranging from silt to boulders. Unlike storm surges individual tsunami waves reach a point of zero water velocity prior to backwash flow. At this point large volumes of sediment may be deposited out of the water column onto the ground surface. Young and Bryant (1992) have made reference to isolated boulders in tsunami deposits in SE Australia. In that area, thicknesses of massive sands and silts include occasional isolated boulders, described by Young and Bryant (1992) as 'boulder floats'. In the absence of any other plausible mechanism of deposition, the processes described above may provide the simplest explanation for their deposition. Isolated boulders contained within massive sandy deposits have been described for other palaeotsunami deposits.

One of the most awkward problems in reconstructing chronologies of former tsunamis for different areas of the world is how to be able to distinguish tsunami deposits from sediments deposited as a result of hurricane-induced storm surges. For example, Liu and Fearn (1993) have shown from coastal Alabama, USA, that a series of hurricanes during historical time have resulted in the deposition of multiple sand layers in low-lying coastal wetlands. Similarly, Davis *et al.* (1989) argued that hurricanes produced graded or homogenous deposits of sand, shell, gravel and mud found in the prominently clastic sediments in the coastal lagoons of Florida. While it is accepted that storm surges result in the deposition of discrete sedimentary units, it is argued that tsunamis in contrast to storm surges, generally result in deposition of sediment sheets, often continuous over relatively wide areas and considerable distances inland. For example, sediment sheets in the Algarve, Portugal associated with the Lisbon earthquake tsunami of 1755 occur in excess of 1 km inland (Hindson *et al.*, 1996 and Hindson and Andrade, 1999; Andrade and Freitas, this chapter). Similarly, palaeotsunami deposits in Scotland associated with the Second Storegga Slide of *circa* 7,100 ^{14}C BP are frequently associated with extensive sediment sheets that occur many hundreds of metres inland (Smith *et al.*, 1985; Dawson *et al.*, 1988; Long *et al.*, 1989). In addition, it may be the case that tsunami deposits contain distinctive microfossil assemblages that can be differentiated from those produced by storm surges (Dawson, 1996).

3.4.2.1 The Second Storegga Slide tsunami

Alastair G. Dawson, David E. Smith, Callum.R. Firth and Shaozhong Shi.

Along the coasts of the northern North Sea, Norwegian sea and north eastern Atlantic ocean a very prominent sand layer, first identified by Sissons and Smith (1965) and originally thought to have been deposited by a storm surge (Smith *et al.*, 1985) has more recently been attributed to a large tsunami circa 7,100 ^{14}C years ago (Smith *et al.*, 1999a). This event was generated in the Norwegian sea as a result of the Second Storrega submarine slide (Bugge, 1983). The widespread deposit is now regarded as a marker horizon against which to compare the age of related deposits and with which to more closely define patterns of land uplift (e.g. Shi, 1995, Smith *et al.*, in press) (see Figure 3.14).

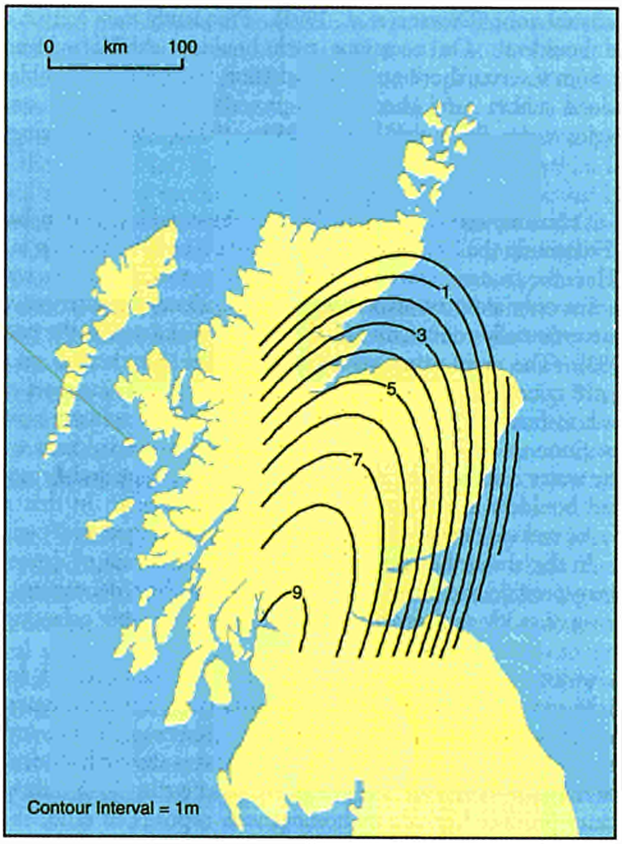

FIGURE 3.15 Isobase map based upon a Quadratic Trend Surface for the Storegga Tsunami shoreline in Scotland, UK (modified following Shi, 1995).

The detail with which the tsunami is known is impressive. Several studies have examined the sedimentology of the deposit, including its particle size (e.g. Shi, 1995), microfossil content and even the time of year it occurred. The diatom ecology of the layer has been examined (e.g. Smith *et al.*, 1992) and over 100 ^{14}C dates on biogenic material both within the layer and from adjacent horizons have been obtained.

Geological investigations of these tsunami deposits on the northern and eastern coastlines of Scotland as well as in uplifted lake basins along the west coast of Norway provide evidence of minimum tsunami runup (Dawson, *et al.*, 1988; Bondevik, 1996; Dawson and Smith, 1997). In eastern Scotland the minimum value of runup associated with this tsunami is in the order of 4-6 metres above contemporary high water mark (Dawson, *et al.*, 1988). However this value, as stated above, should be treated with caution since tsunami flooding to higher elevations may have taken place yet did not leave a sedimentary record. Harbitz (1991, 1992) attempted to develop a numerical model of the Second Storegga submarine slide. He noted that the scale of tsunami runup along the Scottish and Norwegian coastlines very much depended upon the average landslide velocity that was used into the model. For example, he noted that an average slide velocity of 20 m/s resulted in runup values onto adjacent coastlines of between 1-2 m. By contrast a modelled landslide velocity of 50 m/s resulted in runup values of between 5-14 m, values significantly in excess of the estimates for adjacent coastlines based on geological data. Harbitz (1992) concluded that a landslide velocity of 30 m/s provided the closest approximation to the estimated runup values based on geological data. However the weakness in this argument is that the geological data only provide minimum estimates of likely flood runup and therefore the related numerical models of the same tsunami will always underestimate the likely average value of the submarine slide velocity.

The occurrence of this tsunami is unusual since it appears to have been generated by one of the worlds largest submarine sediment slides rather than by an earthquake. This serves to demonstrate that severe tsunamis can be generated by submarine slides in aseismic areas where there are considerable thicknesses of unconsolidated sediments on the sea floor

Giant submarine slides and their potential to generate tsunamis are not restricted to the Norwegian Sea. Recently, Nisbet and Piper (1998) recognised a giant submarine slide occupying the majority of the sea floor of the western Mediterranean. The slide appears to have been generated in deep water adjacent to western Sardinia and radiometric dates appear to indicate that it took place during the last glaciation of the northern Hemisphere (probably *circa* 20-30,000 years BP). At present there is no geological evidence that this submarine slide generated a large tsunami. That such a large tsunami took place, however, can hardly be doubted. However, palaeoenvironmental reconstruction of this event appeared to indicate that the slide took place at a time when regional eustatic sea level in the western Mediterranean may have been at circa -100 metres below present and hence any geological record of the tsunami having taken place may lie below present sea level (Rothwell, *et al.*, 1998). Submarine slides and offshore earthquakes are not mutually exclusive however in their capacity to generate tsunami. For example recent investigations by Perissoratis and Papadopolous (1998) have shown that the islands of Amorgos and Astipalea in the southern Aegean sea were subject to severe tsunami flooding caused by an offshore earthquake in 1956 that simultaneously generated a major sediment slump and the two processes together acted to generate complex patterns of tsunami flooding.

3.5 CONCLUSION

David E. Smith

Extreme coastal floods are a constant reminder that even in the most technologically advanced societies, the importance of natural processes can not be overlooked. Most coastal floods are climate related, and an understanding of their origins, distribution and variations over time may lead to determinations of future trends in their magnitude, frequency and impact. In recent years, progress has been made in understanding the frequency of these phenomena on an annual or decadal scale, but determinations of their impact on coastal processes is at an early stage. The accounts outlined in this chapter illustrate progress towards understanding impacts and trends. In terms of impacts, it is evident that flooding events are considerably variable in their effects, reflecting the many differences in coastal settings. In particular an understanding of rates of change is needed, as Orford *et al.* (1996a) have pointed out. Despite many detailed studies there remains a need for further work before impacts can be characterised. In identifying trends, a number of developments in determining evidence offer hope for the future. Research into the many historical archives of coastal floods offers considerable potential, complementing available instrumental records. Advances in detailed stratigraphical studies will ultimately extend the record, not in the same detail as documentary evidence, but probably in identifying century-scale trends. At the present time, it seems likely that over large areas of the European Union, coastal storminess increased during the Little Ice Age and has probably been increasing since the late nineteenth century.

Tsunamis are less frequent than storm surges, but their effects can be severe, especially in the tectonically active areas of southern Europe and the Mediterranean. It seems unlikely at present that their magnitude and occurrence can be predicted, although progress in compiling records of tsunamis from documentary evidence, identifying tsunami in the stratigraphic record and assessing their magnitude and impact from both the documentary and stratigraphic record offers the prospect that in the future, areas of tsunami occurrence may be more accurately defined and the likely impact of such events assessed. In addition, the effect of tsunamis on the evolution of the coastline may be better understood from work presently being undertaken.

As the secular rise in sea surface levels continues, possibly with a greater increase in high tide levels in some estuarine areas (Jensen, 1993), the impact of coastal flooding will increase. It is possible that thresholds of coastal geomorphic change will be breached locally, but the main impact will be upon the economic and social well-being of coastal communities. There is an increasing need for detailed and relevant information in order to enable effective coastal zone management to be undertaken.

Acknowledgements

The authors would like to acknowledge the work of numerous colleagues involved in the EU projects listed at the beginning of this chapter. Professor J.D. Orford, Dr J.A.G. Cooper and Dr T.A.M. De Groot read sections of the chapter, for which thanks are due. Further, the support of Mr Denis Peter, Dr Anver Ghazi, Dr Ib Troen and Dr M.E. Almeida-Teixeria and Dr G. Fantechi is acknowledged. Ms Erica Milwain drew the diagrams and Ms Gillian West and Mrs Ann Daly typed the manuscript.

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4
CLIMATE CHANGE IMPLICATIONS FOR COASTAL PROCESSES

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4.1 PRESENTATION

Although the topic of this book is Sea Level Change and, theoretically, this chapter should focus on its effects on Coastal Processes, one of the conclusions derived from the Workshop which originated this book is that other climatic change effects e.g. increase in storminess, changes in wind and wave climates could be more important (in forcing a coastal response) than sea level changes.

Due to this, the aim of this chapter is to review the potential effects of climate change on coastal processes and the resulting coastal response. It is structured in terms of "drivers" and "responses" for the sake of presentation, although in nature there is often a co-existence of drivers which act synergetically and lead to a variety of responses.

The chapter has been built from a number of field examples derived mostly from the Mediterranean and North Sea coasts. The conclusions and knowledge derived, however, are intended to be generic and applicable elsewhere.

4.2 DRIVING TERMS

4.2.1 Introduction

Sea level oscillates continuously at a number of time scales varying from seconds (e.g. wind waves) to decades (e.g. sea level change due to global climatic variations). Sea level changes in this chapter will make reference to all changes with a time scale from years to decades or longer, with a comparatively more limited treatment of other Mean Sea Level variations due to e.g. astronomical or meteorological tides (with time scales of hours).

Sea level changes have long been recognised as one of the important driving factors in long-term coastal processes (see e.g. (Sánchez-Arcilla *et al.*, 1996a & 1996b)). Although in the last years the trend has been to focus on sea level rise and its possible acceleration due to anthropogenic factors, it is important to point out that, on a worldwide basis, sea level change is just one factor. In effect, coastal dynamics are a combination of non-linear processes with multiple inputs and outputs interlinked in such a way that the same forcing (e.g. a given sea level change rate) can lead to different coastal responses depending on, for example, the slope and shape of the "receiving" continental shelf and "supplying" watershed basin, the local rates of subsidence, etc. Likewise different combinations of forcings can lead to the same (or very similar) coastal responses (the appearance of so-called "equilibrium" shoreline curves is a possible example). This lack of bi-univocal cause-effect relationship between driving factors and coastal response limits the validity of conventional research projects in which the aim is, for example, to derive or reconstruct a local/regional curve of (relative) sea level rise as the main factor in explaining coastal evolution.

Another common misconception is to consider climate change implications for coastal processes exclusively in terms of sea level change or more precisely, in terms of sea level rise. Any global or regional climate change will certainly affect the relative sea level but also the frequency and magnitude of storminess, the temperature, rainfall and evaporation rates, circulation patterns, etc. This in turn, means changes in the resulting wave-induced transport patterns, in eolian transport, in the availability of sediment, etc. and, of course, in the associated coastal response.

Moreover most of the coasts in developed countries are now evolving under "strong" anthropogenic limitations. River flow regulation, the coastal and river course "rigidization" with buildings and other infrastructures, and even the possible human-induced rise in global temperatures illustrate this point.

To make things even more complex the present knowledge on coastal dynamics, corresponding to a period of relatively steady sea level -e.g. during the last 2,000 years- may not be directly applicable to predict coastal dynamics in a future scenario with an accelerated rate of sea level rise and, possibly, also different rates of sediment supply and coastal configuration. The available numerical/physical models should, thus, be used cautiously to predict coastal behaviour by extrapolating from present conditions.

4.2.2 Driving factors

In spite of recent advances in modelling tools and observational techniques (see e.g. (Sherman and Bauer, 1993); (de Vriend, 1993)) existing observations on driving factors remain sparse in their spatial coverage and limited in their time span, which makes it hard to extract knowledge/conclusions. Likewise the low frequency oscillations of longshore currents, have only been observed and assessed when field data of sufficient duration and spatial coverage have become available (see e.g. Oltman-Shay *et al.*, 1989).

The co-existence of many conceptually different drivers, the different time scales (e.g. from seconds of wind waves to decades of sea level change) as illustrated in Figure 4.1, requires a non trivial methodology to define the main driving terms acting on a given coastal stretch (see e.g. (Sánchez-Arcilla *et al.*, 1995); (Sánchez-Arcilla and Jiménez, 1997); (Jiménez and Sánchez-Arcilla, 1997)).

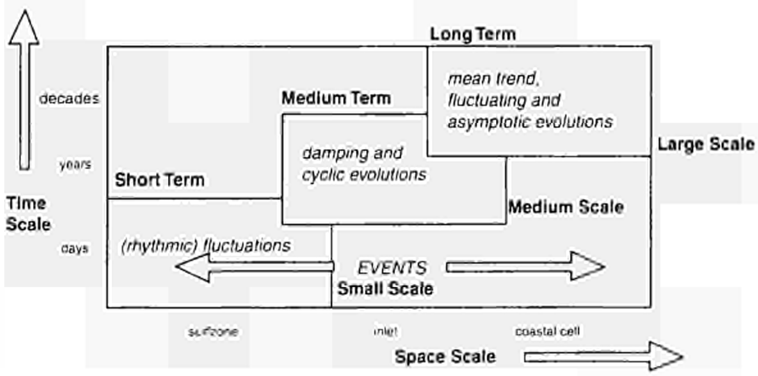


FIGURE 4.1 Illustration of co-existing time-scales in the factors driving the evolution of a coastal stretch (Capobianco et al., 1996b).

All these processes, although combined linearly in some hydrodynamic applications -e.g. linear wave/current or wave/wave superpositions- interact in a non-linear manner when considering the effect on sediment transport (see e.g. (Madsen, 1994)). The combination of waves and currents generates composite bed forms markedly different from those due to waves or current alone (van Rijn, 1993). The co-existence of waves and currents generates a wave boundary layer which modifies the current velocity profile and corresponds to an increased "apparent" roughness, with respect to the pure current case (see e.g. Grant and Madsen, 1986).

Therefore the knowledge of "separate" driving factors may not be enough to predict the resulting joint "driving". This, which limits in general the capability to predict and extrapolate from available necessarily piecemeal knowledge, also applies to processes which are the result of various climatic factors. Swash-zone processes, resulting from the interaction of mean water level, incident waves and the coastal boundary layer of the "underlying" currents, are poorly known and seldom observed or even included in numerical predictions.

This applies in particular to sea level change, the impact of which depends on a complicated interaction with other factors, which limits the predictability and justifies the use of historical, low accuracy reconstructions as a valid analogy for future extrapolations.

Frequently reference will be made throughout the chapter to two contrasting but illustrative coastal environments in Europe: the North West European coast (illustrated by the conditions in and around the North Sea) and the Western Mediterranean coast (illustrated by the coast in and around the Ebro, Rhône and Po deltas). This can help to illustrate the different impact of various effects and processes depending on the type of coastal environment, although all the coasts selected for illustrations are typical of shelf seas environments. The variations correspond to tidal conditions, wind regime and wave climate. Where appropriate, examples from other types of coasts, for example high energy Atlantic coasts, will also be considered.

4.2.3 Climate framework

Global or regional climate change is characterized by changes in circulation patterns, high/low atmospheric pressure systems and variations in temperature and precipitation/evapotranspiration. All these changes will affect driving terms controlling coastal processes such as:

- wind velocity and direction (e.g. aeolian processes)
- other low-frequency climatic events (including mean sea level)
- wave height/period/direction (e.g. wave induced transport)
- circulation patterns (e.g. current induced transport)
- riverine discharges (e.g. continental sediment supplies)

In what follows a detailed description of each of these factors will be presented focusing on their implications for coastal processes.

4.2.4 The wind factor

The coastal and offshore wind climate, characterized by both wind velocities and directions as well as the occurrence of storminess (storms defined in terms of frequency and intensity) is not a constant "driver" but changes with time. Seasonal patterns are commonly quite obvious. But even on a daily scale -breezes- and during storms, the wind velocity and direction are highly variable.

On a scale of decades to centuries information on (coastal) wind climates become less abundant. For the Dutch North Sea coast, the long term changes in coastal wind climate and the potential relation to coastal changes were considered in the early sixties (van Straaten, 1961). This author used paraclimatic data such as the growth direction of trees and the orientation of coastal aeolian landforms to evaluate the effect of changing wind regimes in past centuries. In this area there is a wealth of long-term field observations which allow meaningful analyses for climate change implications. In particular, in the central part of the Netherlands (Amsterdam), wind measurements -basically wind direction- have been available since about 1700 (van Straaten, 1961). The analysis of these series indicates a veering of the wind between 1710 and 1800 (from 235° to 250°) and a backing from 1800 to 1900 (250° to 223°). The latter development is synchronous with the tendency towards a more oceanic climate in the course of the 19th and up into the 20th century (van Straaten, 1961).

This work also shows a striking correlation between the average position of the low tide line, for the central part of the Dutch coast, and the prevailing meteorological/climatological conditions. Severe erosion around 1880 coincided with a period of relatively more frequent W-SW winds (increase of about 3-4%, relative to the period 1843-1866), a decrease in NW winds (of about 3%), much precipitation and a low mean annual temperature. It was assumed that the general changes in climate conditions were due to a greater frequency of storms from the W-SW. In recent literature this period of erosion along the Dutch coast is traditionally referred to as the "van Straaten period" (by Dutch coastal morphologists and engineers).

However in spite of a "statistical" causal relationship between the data, the accuracy of the meteorological information may be rather doubtful. As a matter of fact, long term wind data sets commonly suffer from inhomogeneties due to changes in measuring scales, measuring techniques, the presence of different observers and the effects of a modified environment near the measuring stations (buildings, vegetation, etc.). In some references the conclusion is that wind observations during the last 100 years can not be used to quantitatively analyse changes in storminess (see e.g. (Peterson and Hasse, 1987)) although they are the only existing direct source.

A more recent study of wind and wave data at a decadal scale along the North Sea coast can be derived from observations from light vessels off the Dutch coast (Hoozemans, 1990). For the most important sector of the Dutch coast, the winds appear to be changing over the decades 1960-1980 in comparison to data collected in the first decades of this century (1907-1968). The main conclusions from the analysis of this data are (Hoozemans, 1990): i) a veering of the wind direction, ii) an increase in the average annual wind velocity, and iii) a decreasing percentage of winds from the west sector with consequently a similar increase from the east sector.

Unfortunately the overall trends derived in this work are not fully supported by data from landbased meteorological stations which again casts some doubts on the interpretation of this data. The same applies to other analyses carried out in other European coasts and elsewhere.

In spite of this, even small changes in wind velocity and direction may have considerable consequences for the coastal system in terms of an increase or decrease in water levels. The relation between wind velocity/direction and water levels is particularly relevant for the exceedence frequency of storm surge levels and the risk of coastal flooding. Data from the Southern Bight of the North sea, based on long term tide records (Rijkswaterstaat, 1989) clearly show a relation between wind direction and high water levels (Figure 4.2). For a similar exceedence frequency, a gradual veering of the wind from the West to the Northwest will result in an increase of the high water levels. This effect is mainly a result of the large-scale geometry of the North Sea basin and its orientation relative to the dominant wind direction. Similar conditions may be observed in, for example, the Baltic region or the Mediterranean area, which are very sensitive to meteorological tides. This can also be illustrated by observations from the Northern Adriatic sea, where extreme water levels are generally concurrent with the most severe Sirocco or Southeast storms (Ruol and Tondello, 1997). The particular shape and bathymetry of the Adriatic sea with a longitudinal axis more or less parallel to the Sirocco winds is in favour of this effect. In addition the specific geometry of the Adriatic basin not only affects the meteorological tides but also enhances the astronomical tide. As a result there appears to be a much higher frequency of high water levels in, for example, the Po delta than in other Western Mediterranean deltas, facing a more "open coast" (see e.g. the Rhône and Ebro stormsurge conditions as in Figure 4.4, below from (S.-Arcilla *et al.*, 1998)).

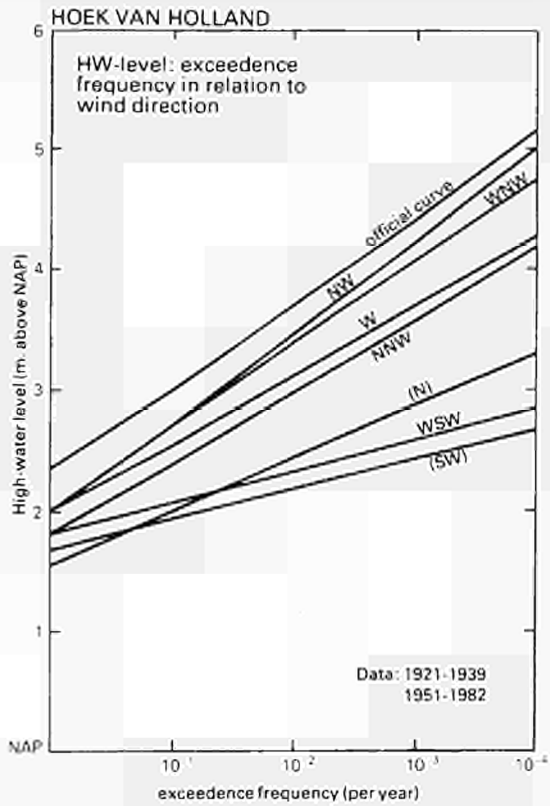


FIGURE 4.2 Relationship between high-water levels (exceedence frequencies) and wind direction in the Dutch coast, from Rijkswaterstaat data.

4.2.5 Other low frequency climate events

Sea level variations on time scales from weeks to decades or even longer are closely related to the internal dynamics of the atmosphere/ocean system. In the analysis of possible trends in observed globally averaged sea level, the high frequency ocean signals, i.e. waves, swell, tides and surges, are relatively easy to remove by filtering (see for example Sturges, 1987). This is however not the case with the low frequency variability since the lowest frequencies that can generally be resolved with the available sea level records are of the same (or larger) amplitude as the globally rising sea level signal (IPCC, 1995). The frequencies are also relatively close which makes the differentiation even more difficult.

On interannual to decadal periods, sea level fluctuations are often driven by atmospheric wind and pressure forcing. Slow variations in wind stress lead to a piling up of water which is mainly modulated by the earth's rotation, friction and the local topography. The atmospheric pressure has an inverse effect on sea level. This is the barometric effect -not considering the so-called geostrophic adjustment- which results in a one cm increase in sea level for one Hpa decrease in atmospheric pressure.

It should be noted that not only the mechanical effect of wind/pressure forcing is important. In addition, the thermodynamic processes and the coupling between atmosphere and ocean must be considered when interpreting long term sea level variations. In particular, the differential ocean heating and changes in salinity may be quite important in connection with climate change. As an example changes in the thermohaline circulation and production rate of North Atlantic deep water may change the northwards transport of upper ocean warm water, which could lead to an enhancement of sea level rise in the mid latitudes of the North Atlantic (Mikolajewicz *et al.*, 1990).

At lower latitudes, for example the tropical Pacific, and at interannual time scales, the El Niño Southern Oscillation (ENSO) coupled atmosphere/ocean phenomenon can clearly be observed in situ -by tide gauges and satellite altimeter observations- and through its effects world wide. During the positive phase sea level is lower than normal at the Western tropical Pacific and during the negative phase (La Niña) it is higher. The opposite situation dominates in the Eastern tropical Pacific (see e.g. (Wyrtki, 1979; 1985); (Miller and Cheney, 1990)). Long homogeneous atmospheric/oceanic observational records which can be used to study air-sea interactions, including sea level, are unfortunately very sparse or non existing. An exception, however, is the ENSO record for the Darwin and Tahiti Stations which covers more than 100 years.

Although the Southern Oscillation has a main periodicity of about 7 years, an increase of positive phases of the event has been detected (in terms of both frequency and magnitude) during the second half of this century. Some authors have associated this increase with the possible influence of climate change (Trenberth and Hoar, 1996) although further data and analyses are necessary to achieve definitive conclusions.

Since these phenomena are the result of atmosphere-ocean interactions at the global scale, their effects are not restricted to the place where they are present but they are "propagated" through the globe via so called "teleconnections". In this sense, El Niño events are accompanied by disturbances in meteorological and oceanographic conditions in the Tropical Atlantic and in the Indian Ocean. As an example, the seasonal latitudinal migration of the ITCZ (Intertropical Convergence Zone) where trade winds converge, is affected by El Niño which produces an early northward displacement resulting in an increase in wind intensity and a decrease in rainfall in NE Brazil (see for example Nobre and Shukla, 1996). Similar teleconnections can also be observed in SE Asia.

Moreover, these oscillations are not restricted to the Southern Hemisphere but they are also observed in the Northern Hemisphere. An example of this is the North Atlantic Oscillation. In summary, low frequency events are present in the ocean and they are propagated through the world ocean by air/sea interactions. Due to this, coupled atmosphere/ocean models become the "natural" tool to study these processes and to forecast their frequency and magnitude through the world. However, although some basic interactions have been successfully modelled, further work is still necessary before this problem can be solved.

In recent years where coupled atmosphere/ocean models have become available, it has however been possible to study many processes in more detail and even discover new phenomena. An example are the air-sea interactions between the subtropical gyre circulation in the North Pacific and the Aleutian low pressure subsystem which has been found in a version of the coupled AOGCM at the Max Planck Institute for Meteorology in Hamburg (Latif and Barnett, 1994). The sea level variations associated with this interdecadal mode are of the order of 3 to 4cm.

For the northwestern European region in recent years there has been much focus on the behaviour of the thermohaline circulation in the North Atlantic, the North Atlantic deep water formation and the Conveyor Belt circulation together with the possible existence of multiple equilibrium states for the global ocean circulation. These are processes, related to anthropogenic/natural climate changes, which will result in regional changes in sea level larger than the global average change, particularly in the North Atlantic region (Micolajewicz *et al.*, 1990).

In addition to these general sea level variations, European coasts are strongly influenced by direct atmospheric forcing. As an illustration, Baltic sea level variations during winter in the present century, can be reasonably modelled using a linear regression model and correlating it with the spatial distribution of monthly Mean Sea Level air pressure over the North Atlantic (Heyen *et al.*, 1996). This so called downscaling model, is often based on canonical correlation analysis which identifies mutual anomaly patterns, in this case between Mean Sea Level air pressure and sea level. Based on these patterns, the empirical model can be used to estimate sea level in months where only the mean air pressure is available, so that it can be used to interpret climate model simulated air pressures in terms of sea level.

The above mentioned model was built with data for the winter months between 1951 and 1970. Figure 4.3 depicts the observed and estimated sea level anomalies for the period between 1899-1987. It can be clearly noticed that there are high correlations between the estimated and observed values, particularly in the northern part of the Baltic. It can be argued that it is the wind stress (which can be related to Mean Sea Level air pressure) which forces the variations in Baltic sea level (Heyen *et al.*, 1996).

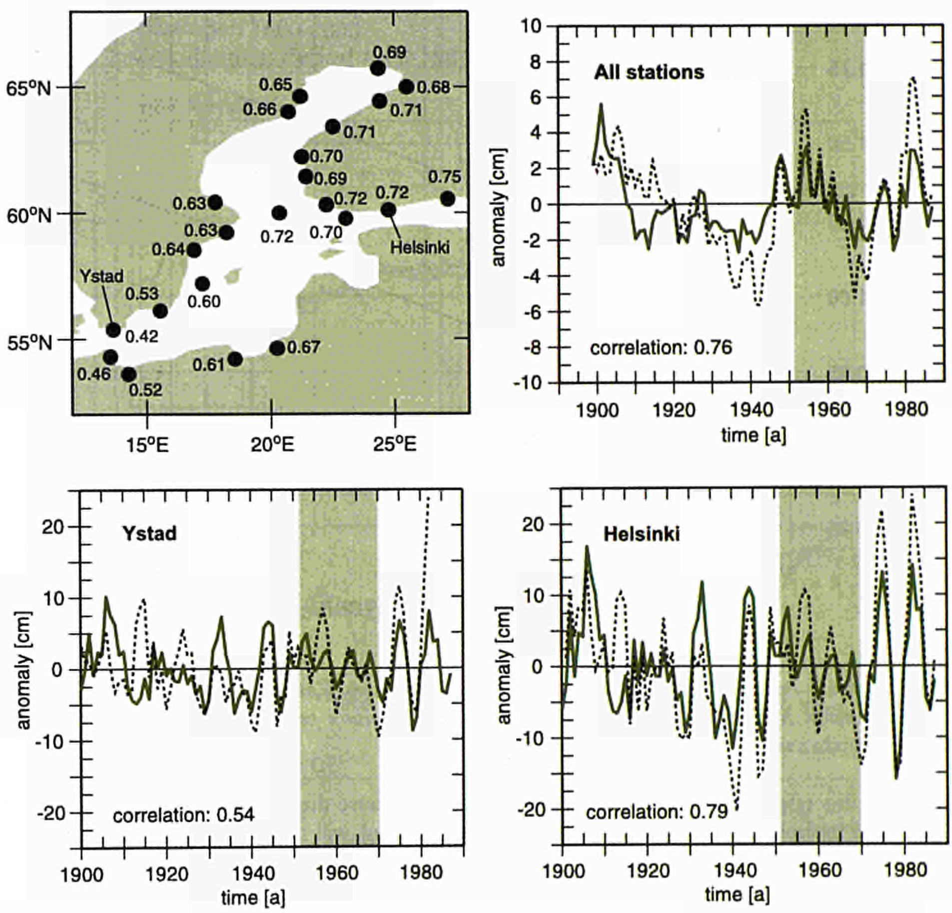


FIGURE 4.3 Observed and estimated (down scaling modelling as explained in text) sea level anomalies in the Baltic Sea -from (Heyen et al., 1996)- for the 1899-1987 period.

In addition to the sea level variations due to meteorological forcing outlined above, water level maxima can also vary in the face of climate change without direct meteorological forcing. For instance, a static coast (that is, one not able to react to relative sea level rise (RSLR)) when subjected to RSLR will experience a decrease in the return period of high water levels due to an increase in Mean Sea Level. Figure 4.4 shows the storm-surge climate for 3 low-lying areas of the Mediterranean Sea, the Ebro, Rhône and Po deltas, under present conditions. If an increase in Mean Sea Level of 25 cm. for the year 2050 is assumed, the present high water level associated to a return period of 50 years will have a corresponding return period of 13 years for the Ebro delta, 8 years for the Rhône delta and 4 years for the Po delta (Jiménez and S.-Arcilla, 1997); (Ruol and Tondello, 1997). As it can be clearly seen, for relatively "mild slope" water level curves, this effect will be more significant, whereas for relatively "steep curves" as e.g. the North Sea, the decrease in return periods will be less important (see e.g. (Delft Hydraulics, 1993)).

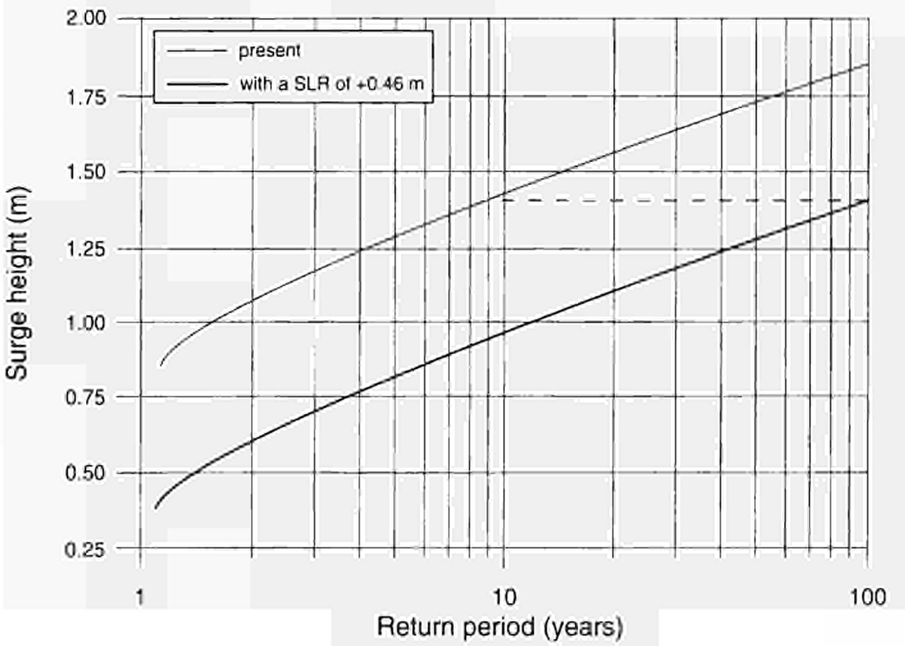


FIGURE 4.4 Probabilistic distribution of storm surges in the Ebro, Rhône and Po deltas -as obtained in the EU-Environment MEDDELT project- showing the decrease in return periods for a given increase in MWL (Sánchez-Arcilla and Jiménez, 1997).

Although this effect will not likely affect active coasts, because they will re-adapt to RSLR, passive coasts and monolithic ones (as e.g. those rigidized by man) will certainly experience it (S.-Arcilla and Jiménez, 1997).

This is an example of the synergic action between different factors related to climate change (CC) variations, where it is emphasized that some global changes due to climate change (e.g. eustatic sea level rise) will have different effects world-wide according to the local or regional climate characteristics (in the case of the example depending on the regional storm surge climate).

This is further demonstrated by recent data on mean relative sea level rise. The present trend of relative sea level rise for the Mediterranean, based on the stations with the longest records, is less than 2.2 mm per year (Tsimplis and Spencer, 1997). For the Adriatic coast, however, there appears to be a constant and coherent trend since 1950 (i.e. a trend without rate of increase). The relatively short return periods for the high water levels near the Po delta again emphasizes the effect of local meteorological tides, in this case in combination with local astronomical tides -since the northern Adriatic is a remarkably good reflector of the incident M2 tidal energy (Hendershott and Speranza, 1971).

Tidal records from stations along the North Sea coast of the Netherlands indicate a relative sea level rise (based on average tidal levels) varying from 11 to 22 cm per century. The longshore variation is mainly due to differential vertical movements (subsidence) along the coast (Lorentz *et al.*, 1992). Average tidal levels from the mouth of the Westerschelde in the Southwest of the Netherlands, have increased about 30cm in the last 150 years. In this particular location, the mean high water levels rise faster than the Mean Sea Level. The mean low waters, in contrast, show a much lower rise than the average tidal level. As a consequence tidal ranges have increased in the estuary (Figure 4.5).

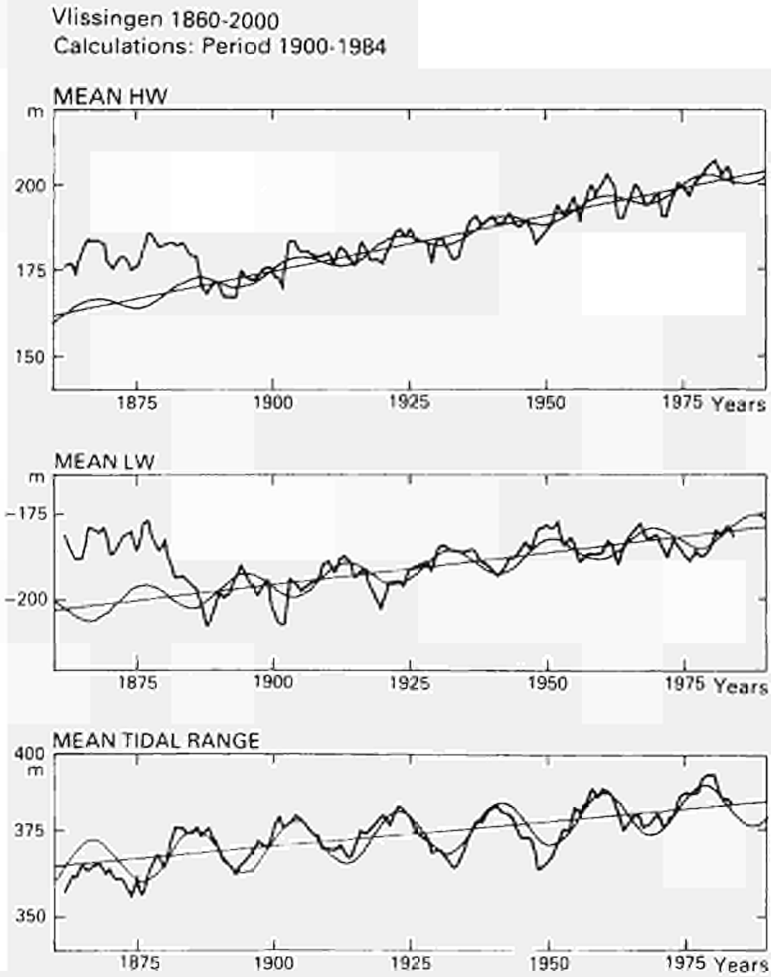


FIGURE 4.5 Time evolution of mean high-water and Mean Sea Levels in the SW Dutch coast (Source: Rijkswaterstaat).

Apart from an average trend, the curve for the southwestern Dutch coast is also dominated by truly periodical oscillations with a period of 18.5-19 years. This corresponds to larger astronomical periodicities than in the tidal cycle (in this case the period of 18.6 years has been related to the rotation of the lunar orbital plane (Pugh, 1987; Gerritsen, 1986). The amplification of the tide in the Westerschelde estuary is explained by a number of new "developments" (Van den Berg *et al.,* 1996). The disappearance of side branches and the reduction of intertidal areas along the shore lines has resulted in a more funnel-shaped estuary, in combination with an increase in average depth. The corresponding decrease in frictional damping and increase in contraction of tidal energy leads to an increase in tidal amplitudes. The increase in average depth can also very likely result in a better resonance of the tidal wave and therefore in an associated increase in tidal range. The dredging activities in the estuary and the artificial deepening of channel bars has produced a significant reduction in hydraulic resistance which in turn have induced a lowering of the low water levels. Finally there appears to be an amplification of the tidal waves in the North Sea basin, which causes a small increase in tidal amplitudes along the coast (Misdorp *et al.,* 1990). For the northern part of the Dutch coast the increase in water levels is sometimes explained by a northward displacement of the amphidromic point.

In summary an increase in Mean Sea Level may be accompanied by a change in tidal amplitudes in coastal regions due to changes in frictional damping and in the configuration of coastal bays, tidal basins and estuaries. The associated increase in tidal current velocities will result in higher transport rates, in both flood and ebb conditions. The long term consequences of this effect heavily depend on the potential feedbacks and responses in the system (e.g. wave sediment availability, incident wave storms, etc.). In particular the degree of tidal asymmetry and the basin geometry are considered to be quite relevant (van der Berg *et al.,* 1996). Tidal asymmetry favours a net import of sediment. This process gradually slows down as sandy shoals start to emerge. The presence of shoals directly implies a decrease of flood current velocities and an increase in ebb velocities, reducing thus the effect of tidal asymmetry. Simultaneously, a decrease in average water depths and the corresponding increase in frictional damping reduces the tidal amplitude again. Therefore it is the combination of locally existing factors -tidal asymmetry, basin geometry, sediment availability and anthropogenic factors- which determine the actual tidal regime in response to sea level rise.

4.2.6 The wind waves

There is, in general, a strong coupling between wind and wave conditions, especially in coastal areas subjected to locally generated wind waves. For swell dominated coasts, however, such a relation may be less obvious and harder to link to the more local climate changes. Changes in wind velocity and direction will affect wave growth (both wave height and period), modifying the total wind stress at the water surface, by limiting or increasing the potential fetch length and by determining the duration of the wind field. A change in any of these individual parameters will locally modify the wave climate.

Long term records of wave data may reflect changes in wave climate but the interpretation of these data sets is again limited by the presence of inhomogeneities of various types. Again, the North Sea behaves as the best data source for this analysis because it possesses the longest records. The available data sets seem to indicate a "recent" increase in significant wave height in the North Atlantic (see for example Carter and Draper, 1988). A similar tendency has been reported for the Dutch North Sea coast, for the period 1960-1980 (Hoozemans, 1990). Visual observations from three Dutch light vessels, in combination with recorded wave heights by a British vessel, independently show an increasing trend in the significant wave height (Hoozemans, 1990) (Figure 4.6).

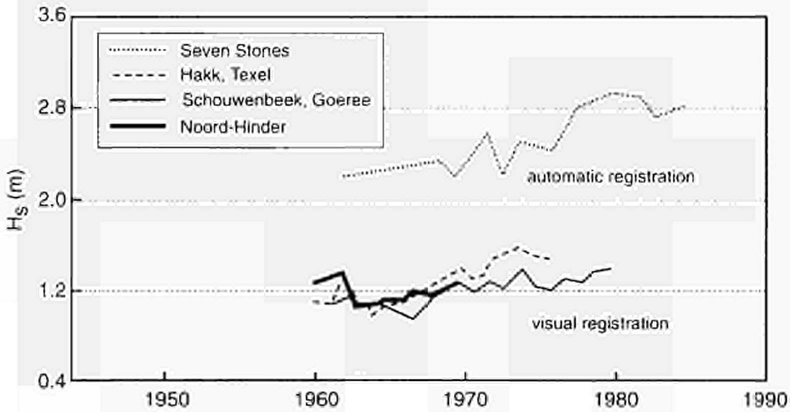


FIGURE 4.6 Decadal scale evolution in significant wave height -in terms of mean annual Hs- from visual Dutch data and registrered British data (Hoozemans, 1990).

Predictions of changes in future wave climates could also be based on accurate wave generation and propagation models fed by the predicted atmospheric climate. Since wave generation models are normally considered to be more accurate than meteorological models (Koman *et al.*, 1994) most of the uncertainties should lie in the "atmospheric component".

As expected the wave climate also tends to have a seasonal component. This can be illustrated by wave measurements off the Ebro delta in the North-western Mediterranean which show a well defined seasonal pattern (Jiménez *et al.*, 1997a and 1997b). Relatively low waves (Hs ca 0.6m) incident from a Southern direction are typical for the period from June to September (Figure 4.7). The most energetic waves (Hs ca 0.9m) can be observed from October to March and are incident from the E-NE or the NW. Storm conditions will occur for all three wind directions although the directional distribution is not equally spread in time. The directional distribution of recorded storms in the period 1990-1994 is characterized by Northwest storms for 47.8% of the observations and East and South storms for 38.9% and 13.3% of the observations, respectively. Eastern storms result in slightly higher significant wave heights than for other storm directions. Strong Easterly winds are commonly associated with the frequent occurrence of a centre of low pressure located over the Mediterranean seaward of the Spanish coast (Jiménez *et al.*, 1997a, 1997b and 1997c). These strong easterly winds also produde significant meteorological tides along the coast. The corresponding average monthly surge height is about 0.1-0.2m, with maximum monthly surge heights in the range of 0.3-1.0m (Jiménez *et al.*, *ibid.*). These meteorological tides and related surge heights are commonly more important for coastal processes than the astronomical tides in microtidal environments.

The conditions in the Ebro delta previously described demonstrate the existence of coupled hydro-meteorological scenarios. Developments in wind conditions, wave fields and water levels (surge levels) often show strong coherence. An increase in wind velocities, during storms, results in a growth of wave height, specially in the case of "optimal" fetch and will also result in substantial surge levels along the coast. From a coastal-morphodynamic point of view, the combination of storm waves and surge levels is expected to result in a strong morphological response of the coast.

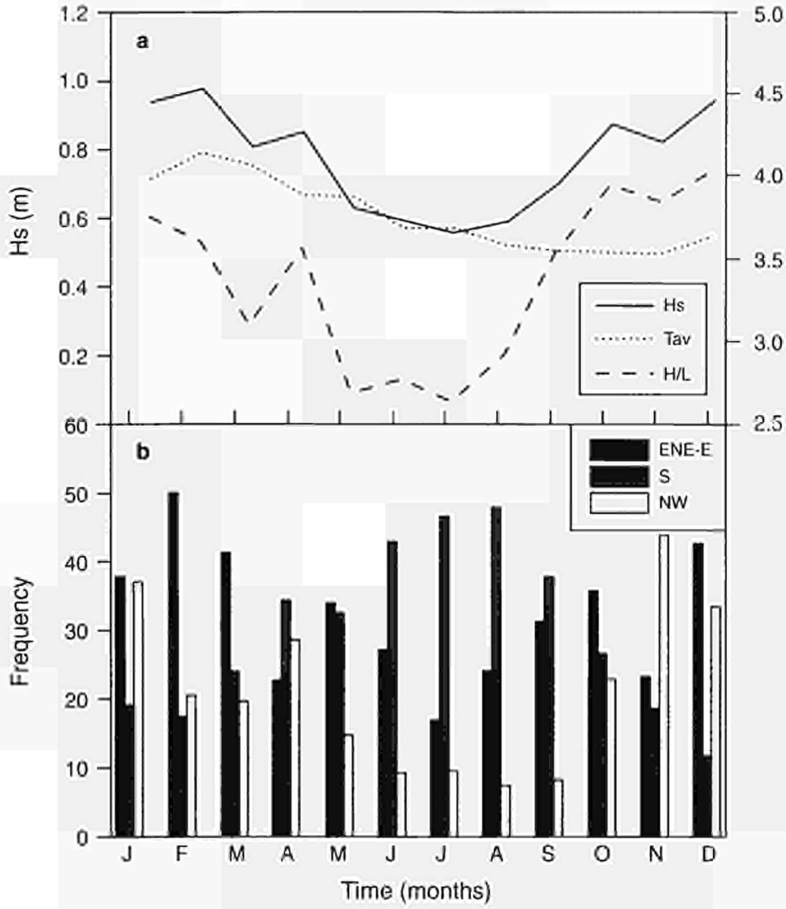


FIGURE 4.7 *Seasonal distribution of offshore wave parameters at the Ebro delta coast (a) Hs and Tm and Hs/Tm; b) wave direction) (Jiménez et al., 1996).*

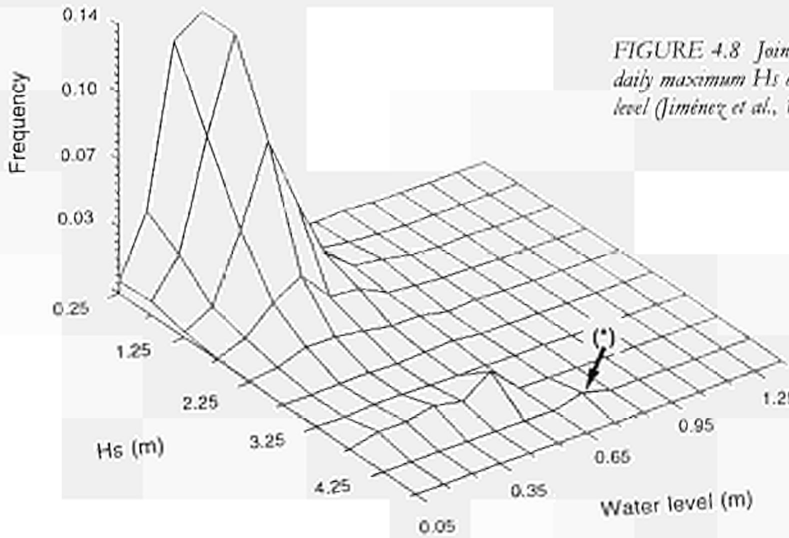


FIGURE 4.8 *Joint presentation of daily maximum Hs and mean-water-level (Jiménez et al., 1996).*

For the Ebro delta coast, the mutual relationship between wave height and surge levels is illustrated in Figure 4.8. In this example there appears to be no strong correlation between waves and water levels since the highest waves do not often coincide with the highest water levels. Along the Southern Bight of the North Sea, however, the conditions are markedly different (Figure 4.9). The expected wave height and storm surge levels show a strong correlation together with an increase of storm surge levels relative to Mean Sea Level -accompanied by a further increase in the expected wave height (RijksWaterstaat, 1990).

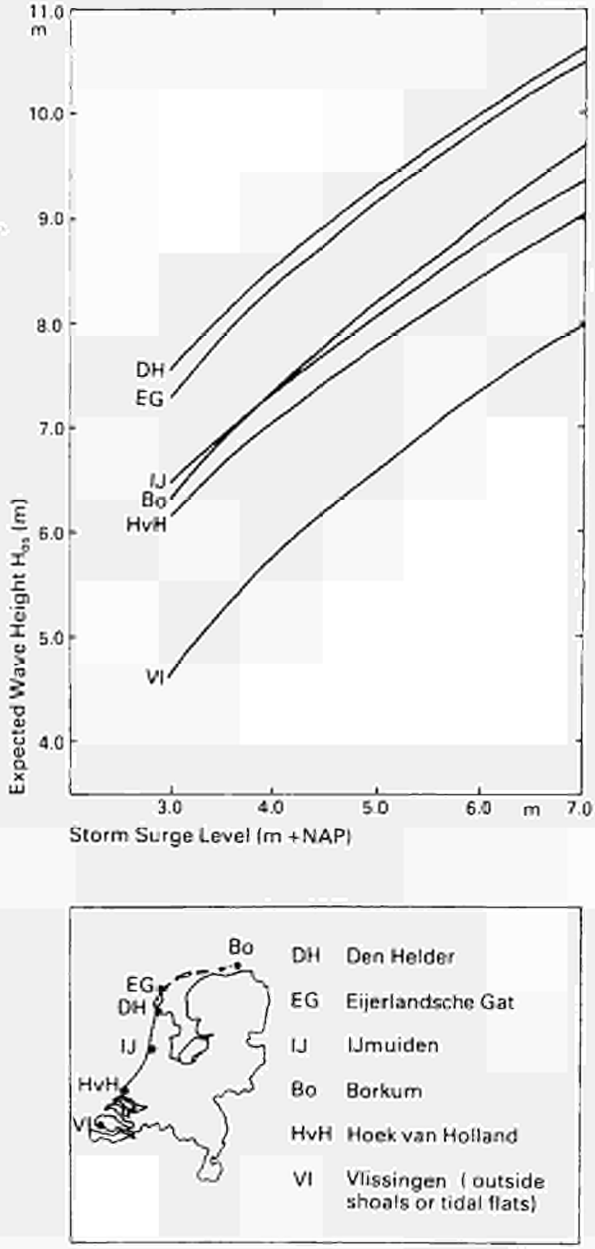


FIGURE 4.9 Correlation between expected wave heights (Hs in deep water) and storm surges (Rijkswaterstaat, 1990).

Therefore, the impact of a future change in wind/wave intensity and direction may vary from region to region. The North sea case is clearly different from the Mediterranean (Ebro delta) case. The decoupling of various effects again stresses the fact that a simple and uniform extrapolation of conditions is not possible.

Because of the differences in time scales, changes in sea level do not result in a sudden change of water depth and associated nearshore wave conditions. Therefore there is not a clear causal relation between Mean Sea Level changes and wave height conditions other than the ones previously described for storm surge levels (coupled scenarios). In the highly dynamic nearshore zone, the gradual change in water level tends to be compensated by an equivalent gradual change in beach profiles (for instance in the form of a tendency to have a profile migration landwards as given by Bruun's rule).

Conditions may be markedly different for areas experiencing a much more rapid rise in relative sea level such as surrounding the Mississippi delta. Tide gauge records from Eugene Island show a relative sea level rise in the order of 1.6cm per year (Coleman and Roberts, 1989). The lack of sediments together with the existing subsidence and the increase in water depth in the seaward part of the cross-shore profile lead to a substantial reduction in cross-shore wave energy dissipation and therefore to an increased amount of nearshore wave attack. Near the shore line, the more energetic waves are responsible for stronger concentration of wave breaking and a more intense development of wave driven currents. The beaches will be characterized by stronger swash zone processes, overwash processes and breaching (S.-Arcilla and Jiménez, 1994; McBride *et al.*, 1995).

4.2.7 The nearshore currents

Shelf currents are mainly driven by tides (and other mean water level variations), winds and thermohaline factors. In the nearshore zone, the wind role is enhanced but the main driving factor is usually the incident wind waves. The wave induced circulation pattern in the nearshore zone is strongly 3D and interactive with the coastal shape and bathymetry (Sánchez-Arcilla and Lemos, 1990).

The resulting circulation will change with the existing sea level, since sea level changes modify the geometry of e.g. the nearshore domain (sea level rise translates the nearshore and surf zones shorewards). The resulting currents will also vary with the changes in atmospheric climate (e.g. wind distribution) and marine climate (e.g. wave conditions).

The impact of changes in the general circulation pattern or even the mesoscale features of the shelf currents is likely to be less "intense" since the corresponding factors are much less energetic (Sánchez-Arcilla *et al.*, 1998). This means that the morphodynamic impact will also be less significant.

In the nearshore area different types of circulation can be distinguished in both the horizontal and vertical plane. The cell-like circulation systems develop according to a variety of temporal and spatial scales and their presence may have a daily, seasonal or annual pattern. These flow patterns may be restricted to the zone of breaking waves (surf zone), enlarged to cover the full nearshore area, and can also be extended to the shore face and/or shelf domains.

The wave driven littoral drift, frequently associated with the development of littoral cells due to an irregularly shaped coastline, is a well known feature along wave and wind dominated coasts. The large littoral cells and the existing longshore sediment transport patterns, result from a subtle balance between coast geometry (beach profile and shoreline orientation) and wind plus wave characteristics (angle of approach, energy level, etc.). The resulting circulation is thus clearly dependent on any change in wave/wind characteristics due to a local climate change (e.g. longshore current velocities vary roughly with the sine of the angle between coastline and wave crest orientation).

The effect of sea level rise on wave induced circulation is therefore threefold:

i. A change in mean water level will affect the wave propagation features and in particular, the position of the breaker area.
ii. A change in mean water level will modify the coastline configuration and bathymetry of the domain where the currents flow and, thus, the resulting current velocities.
iii. A change in mean water level will, thus, also modify the turbulence level and, correspondingly, the resulting flow pattern.

Nearshore circulation and sediment transport rates are also expected to be affected by changes in wind velocity and direction since wind driven flows can play a dominant role in shallow water conditions. From the analysis of nearshore current and wind data at the Dutch site, it was observed that for similar wave conditions (wave height and direction) the longshore current could vary by a factor of three, depending on the strength of the wind (Hubertz, 1986). Determination of the role of the surface wind stress and its contribution to the longshore current was theoretically based on wind and wave forcing terms in the longshore momentum equation (Whitford and Thornton, 1993).

By comparing the total wave and wind forces for increasing wind speeds at constant angles (between wind and wave directions) it was noticed that the wind force increases more rapidly than the wave force for increasing wind speeds. From surf zone measurements in Terschelling (The Netherlands), similar trends are found for wave and wind driven flows. Even for moderate wind conditions (smaller than 8 in the Beaufort scale) wind driven flows have the same order of magnitude as tidal currents, near 0.35-0.40 m/s (Hoekstra *et al*, 1994). For stronger winds, the wind driven longshore flow in combination with the wave driven flow tend to dominate the nearshore flow regime.

In a cross-shore direction, wave and wind driven flows are present in the form of rip currents, of oblique oriented "feeder currents" and undertows. Rip currents are part of the local cell circulation system which compensates the onshore mass fluxes due to the presence of shoaling and breaking waves. These rip currents are fed by "feeder currents", which develop in response to longshore inhomogeneities (in the wave field for example or as a response to the nearshore bathymetry). A compensation of mass fluxes is also the main driving force for undertow. The wave driven undertow is commonly enhanced by the effect of a seaward sloping water surface generated by on-shore directed wind stress. Both effects (wave and wind driven undertows) are in practice hard to separate. The origin, development and dynamics of rip currents are only partly understood and although they are typically created in shallow depths it is difficult to assess the quantitative effect of sea level rise or climate change except in terms of scenarios which contain a number of "educated guesses". Any eventual change in mean water level will modify the geometry of the domain, and correspondingly, the resulting fluxes. For instance, an increase in wave height will extend, seaward, the surf zone and the magnitude of the longshore current, feeder currents and rips and undertow. A change in wave period or wave direction will have less straightforward effects.

It is therefore difficult to assess quantitatively the final effect of this scenario type of change which may result in a more complicated interaction of the various longshore and cross-shore flow patterns.

4.2.8 The shelf currents

Shelf sea environments are typically restricted to depths greater than 20-25m and reach a maximum depth of about 200m. The shelf generally slopes in the order of 1:1,000. The lower shoreface covers roughly an area between the 10 and 20m depth and forms the link between the upper shoreface or nearshore zone and the shelf. In this area the coastal slope varies from about 1:100 to 1:500.

Shelf sea circulation patterns, as mentioned in the previous section, are the result of astronomical tides, wind and density driven effects, and the interaction of flows with coastal morphology and bottom topography. Given the considerable range of depths, a gradual change in sea level will only be of minor importance for shelf sea circulation patterns. For instance a 0.5m increase in mean water level is a 0.5% increase for the 100m isobath and compares well with the error associated with many bathymetric data sets.

Shelf sea circulation patterns in the North Sea can be used to further illustrate the potential impact of change in wind conditions and density patterns (Figure 4.10). Water masses from the North Atlantic and the Strait of Dover enter the North Sea and generally mix with fresh river water and low salinity water from the Baltic. Inflow from the Atlantic is due to the predominantly westerly winds, the propagation of the tidal wave and density differences (Eisma, 1987). The mass flux through the Strait of Dover is essentially the result of the residual effect associated with tidal interaction, a wind driven component and a long term gradient in sea level of the North Sea. The average annual transport by all three factors is estimated to be of about $150 \cdot 10^3 m^3/s$ (Groenendijk, 1988). This effect could alone be responsible for an average residual current over the shelf in the Dutch coastal zone (depth between 25-40m, width between 100-150km) of about 3-7cm/s (de Ruijter *et al.*, 1987).

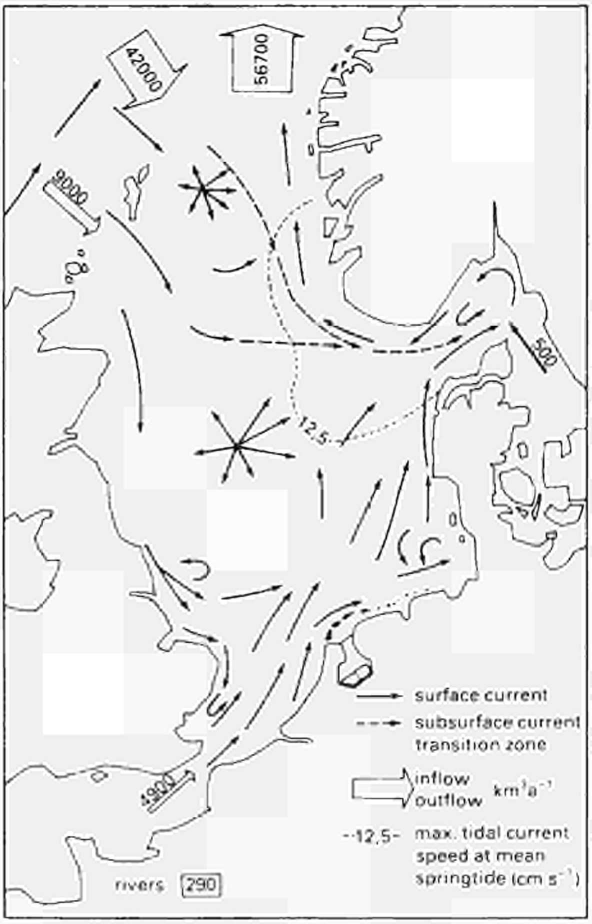


FIGURE 4.10
Schematization of the North Sea basin circulation pattern (Eisma and Kalf, 1987).

The general residual circulation in the North Sea is characterized by a more or less southward flow along the English coast to the Southern Bight of the North Sea. From the Southern Bight and the Strait of Dover residual currents promote a northward flux along the Belgian, Dutch, German and Danish coasts to the Skagerrak (Figure 4.10.) (Eisma and Kalf, 1987). However along the entire West coast of continental Europe flow patterns and residual currents are easily affected by local wind fields and density effects. A change in wind force will alter the general North Sea circulation and, in addition, the wind factor can completely reverse the normal flow pattern at more regional scales. For example a steady Western wind gives a large West-East residual flux in the North sea and a piling up of water in the German Bight. Afterwards, the strong westward flow dominates for some time (Gerritsen, 1986), thus releasing the water from the German Bight.

Along the Dutch and German coasts, the north and northwest-directed residual flux of water also becomes enriched with increasing amounts of fresh water from a number of rivers. The largest contribution of fresh water is from the Rhine, with an annual discharge varying in the range 800-10,000m^3/s (with an average of about 2,500m^3/s).

The combination of wind stresses and density gradients is particularly important for nearshore and shelf sea circulations. Horizontal density gradients such as those depictured in Figure 4.11 -based on measurements along the central Dutch coast- are responsible for density driven cross-shore circulation with surface flow in an offshore direction and a consistent shoreward directed velocity component near the bottom (Figure 4.12) (van de Giessen *et al.*, 1990). In this case a very strong shoreward-directed residual near bottom flow (about 7cm/s) is observed in conditions with ENE-SE winds, when wind and density effects enhance each other.

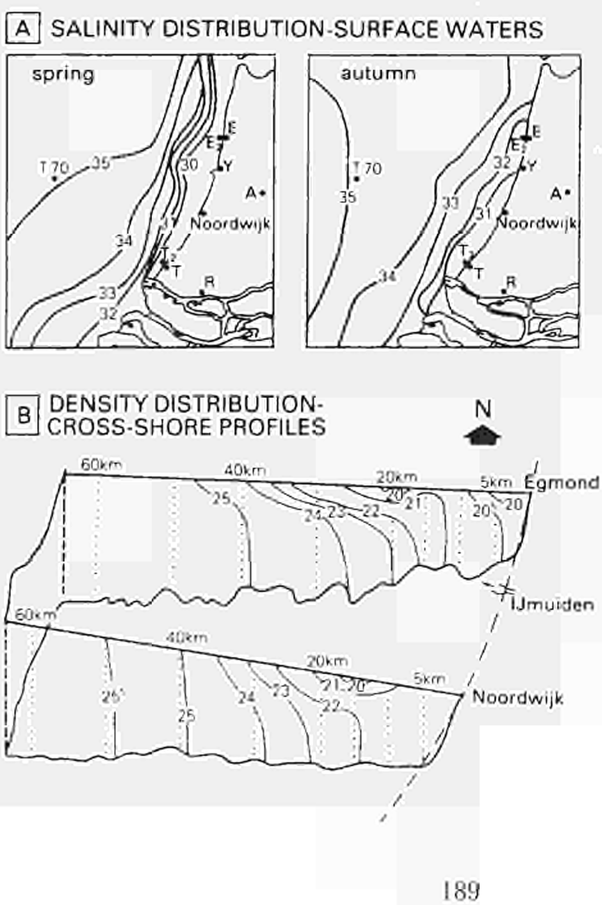


FIGURE 4.11 Horizontal density gradients near the Dutch coast, as driving factors for the density-driven cross-shore circulation (van de Giessen et al., 1990).

In extremely stratified conditions, however, associated with high discharge events from the Rhine river (e.g. exceeding 6,000m³/s) a decoupling occurs between the upper layer of relatively fresh water and the high density layers of saline water near the bottom. Consequently strong winds have only a minor impact on the bottom currents and, as a result the cross-shore residual current mainly responds to the lateral density gradients (van de Giessen *et al.*, 1990). The passage of the core of a large fresh water plume, occasionally results in an onshore directed residual bottom flow of about 10cm/s. Although most observations are based on measurements in the so called Noordwijk transect (Figures 11 and 12) the conclusion is clearly that the density gradients have far reaching consequences for the magnitude of the residual bottom flow.

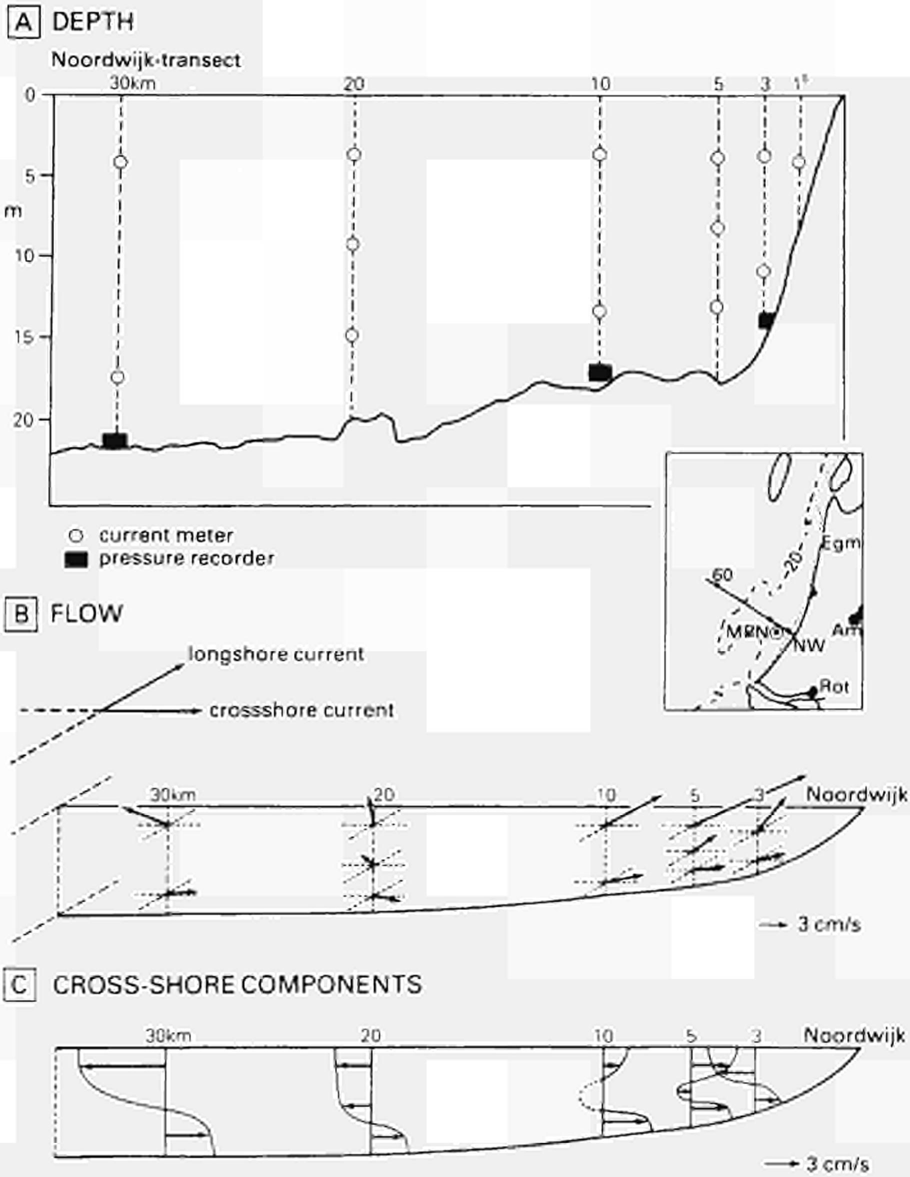


FIGURE 4.12 Cross-shore circulation pattern corresponding to the case shown in Figure 11 (from de Giessen et al., 1990).

At the shelf border, the slope-jet current is well developed in most continental shelves and shows also important variations with seasonality and climate changes (Figure 4.13). In the Ebro delta region, as an illustration, it is easy to see that in the winter season the front is well established with geostrophic currents of order 50cm/s. In autumn the front is less marked showing in the upper layer (0-100m) unstable patterns with evidence of cyclonic and anticyclonic eddies travelling along the front. Finally in the summer season there appears to be a reversal of the current in the narrower (northern) region, while the flow in the wider (southern) region remains anticyclonic. This is due to an offshore anticyclonic eddy which appears as a result of the balance between the slope jet, the wind field and the river discharges.

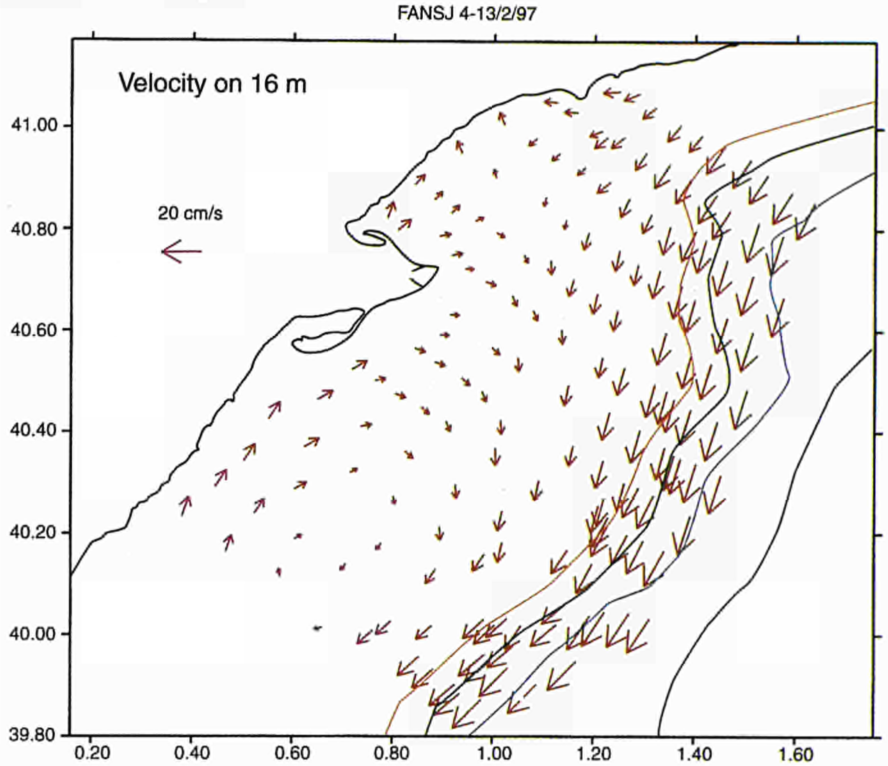


FIGURE 4.13 Slope-jet (seasonal) currents in the Ebro delta region as observed in the MAST-III FANS project (Sánchez-Arcilla and García, 1997).

This clearly illustrates that any change in the atmospheric or marine climate, which would affect the general circulation in the area (e.g. the slope jet), the wind field in the area (e.g. surface current and coastal mean water levels) and the river discharges, would significantly affect the transport patterns and therefore the associated coastal processes.

Likewise the exchanges of water between shelf and slope, which also affect the exchanges of sediment and nutrients and therefore the underlying coastal processes, show also a high variability with seasonality and thus climatic conditions. As an illustration, in the Ebro delta area, the slope water is usually imported via near bottom intrusions, while the shelf water is exported via surface filaments or mesoscale currents at intermediate depths. These exchanges depend (FANS-Blanes, 1997; FANS-Bangor, 1998) on riverine discharges and mesoscale circulation features, both of which are clearly sensitive to climate changes.

In summary the direct sea level rise effects are expected to be of minor importance for shelf sea circulation and more dominant effects may be expected from changes in the regional wind conditions or the run-off regime of major rivers.

However, at a larger scale, the aggregated effects of sea level rise play a more dominant role. The circulation patterns in the Mediterranean from the Miocene to the present day (Pujol, 1989); (Thunell and Williams, 1989) may have controlled the Mean Sea Levels and also the longshore progradation of e.g. deltaic lobes across the shelf and even to some extent the dynamics of sandy beaches (see e.g. (Díaz *et al.*, 1990)). During the course of the main Holocene transgression, the changes experienced in sea level controlled the landward displacement of the coastal environment and the development of lag deposits and erosional surfaces on the shelf (Figure 4.14).

4.2.9 The river discharges

Riverine supplies are one of the sources of sediment for the coastal zone. This implies that any change in river regime, even assuming steadiness in marine climate, will seriously affect coastal evolution. While this factor will affect all the sedimentary coastal systems, deltas in particular and low-lying coasts, in general, will be the more sensitive environments for such changes.

In opposition to other changes analysed in driving forces, changes in river regimes are produced inland and they propagate towards the coast using the river as a conductive vector. Within the river basin, changes in rainfall, evapotranspiration and vegetation will induce different run-off regimes and correspondingly changes in the fluvial liquid and solid discharges. For the receiving coastal system these changes in upstream boundary conditions can be translated into an alteration of the outflow regime and associated sediment dispersion. The adaptation of the coastal system associated to these changes in the input of water, sediment and nutrients will result in modifications.

The studies on the influence of climate change in river discharges and, in particular, at an "aggregated" level (which would correspond to the entire river basin) are relatively recent (see e.g. (Day *et al.*, 1995)). For the Rhine river, the Rijkswaterstaat, based on the ISOS (Impact of Sea Level Rise on Society) study assumed a maximum increase in winter discharge of 10% and a maximum reduction in summer discharge of also 10%. These figures were based on the result of general circulation models, that indicated that there would be a global increase of 10% in both precipitation and evapotranspiration. In addition, these assumptions were also based on the expectation that, due to an increase in temperature, there would also be an increase in winter snow melt in the mountainious part of the river basin.

More sophisticated studies for the Rhine river have used a spatially distributed water balance model which simulates monthly discharges from the main river and its major distributeries using meteorological data (Kwadijk, 1993). By realizing a coupling with an "emissions model", a climate model and a general circulation model, it was possible to determine some future trends in river discharge based on the business as usual (BAU) and the accelerated policies (AP) scenarios of IPCC (Kwadijk, 1993). The results show that there appears to be a significant range of uncertainty which is mainly due to the wide range in the simulated precipitation results derived from the general circulation models. As a general conclusion the winter discharges will increase while the summer discharges will decrease and the present combined snowmelt/rain-fed river regime will change into a dominantly rain-fed river regime. Besides the present existing seasonal contrast in river discharges -with largest average discharges in late winter and smaller discharges in autumn- will be further enhanced.

This type of model is not capable of predicting changes in the frequency of extreme flood events. However, this type of model can still be used in a more statistical sense combined with the available

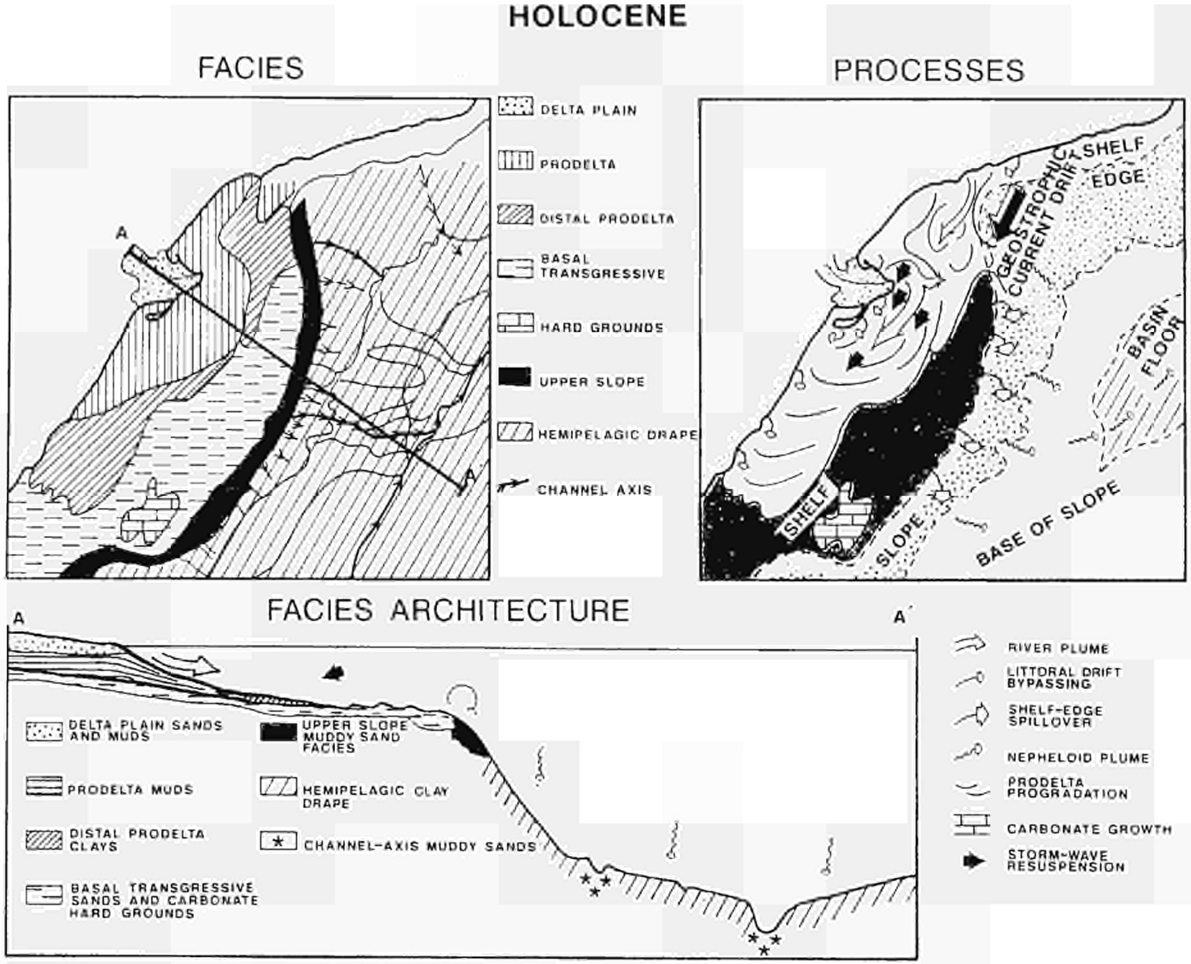


FIGURE 4.14 Generalized sedimentary facies and processes of the Holocene sea level highstand showing distribution of depocentres and growth patterns of the Ebro coast and shelf, after present sea level was reached (Nelson and Maldonado, 1990).

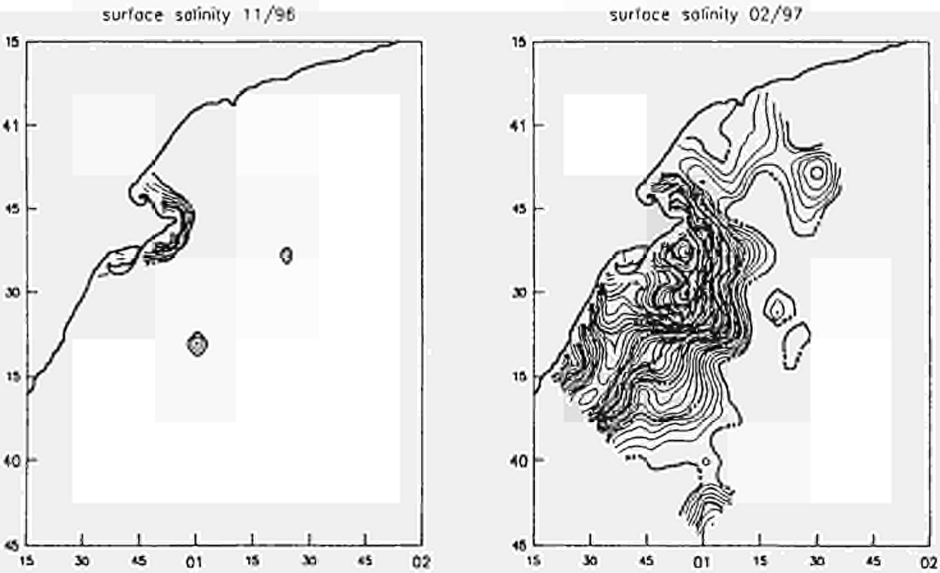
data to estimate changes in recurrence intervals and height of peak floods (Kwadijk, 1993). With this approach, it was found that flood frequencies are more sensitive to a change in precipitation than to a change in temperature. The present two year return period peak flow for the Rhine river is of about 6.500-7.000 m^3/s. With a precipitation increase of 20%, this peak flow will be raised by about 30%, while a reduction in precipitation of 20% will lead to a lowering of the peak flow by about 30%. A temperature rise of 4° C will lower the peak flow by about 6% (Kwadijk, 1993). At a more "local" scale, the river discharge influences the sea surface temperature and salinity and, therefore, the biological productivity in the area and the corresponding distribution of sediment and nutrients. Taking the Ebro river as an illustration (Salat *et al.*, 1998), for river discharges of the order of 10 times the average discharge, i.e. about 2,000m^3/s, the surface temperature is significantly lower than the expected average (about 12° for the winter season of 1997) and the sea surface salinity also shows this effect (36.5 psu).

Depending on the degree of mixing, which is related to the climate situation in the area, the low salinity fresh water may spread in various manners. When the depth of the surface mixed water is limited (e.g. at the summer season when this layer is of the order of 50m) the stirring of low salinity water is enhanced because of mass conservation arguments (Figure 4.15). The resulting plume dynamics are therefore strongly linked to and dependent on meteorological/oceanic conditions and any climate change even of minor magnitude.

To go one step further, i.e. from water discharges to sediment/nutrient discharges, it is necessary to deal with increased uncertainties since many other variables (both from the natural and socio-economic systems within the river basin) come into play. For instance a change in precipitation and evapotranspiration will not only affect the discharge regime but will also initiate changes in natural vegetation, land uses and degree of soil erosion. Soil erosion itself is already a complex process and even for present conditions, sediment delivery to the river system is often highly variable and difficult to compute. Therefore the estimates of sediment discharge and response to future climate changes are, in general, beyond the capability of presently existing models.

These variations in liquid and solid discharges may be quite irrelevant when considering the impact of anthropogenic effects in the river basin and associated river deltas during the last decades. The increasing demand for a better water management during the last century has seriously affected many deltaic systems (see e.g. (Stanley and Warne, 1993)). Riverine sediments are trapped by dams and in reservoirs, high discharge events have almost disappeared -due to increased storage capacity- and natural flooding and deposition in coastal wetlands has been greatly reduced as a result of the previous effects (in combination with the implementation of more efficient irrigation networks and increasing implementation of channels and dykes).

In deltaic areas, this decrease in sediment supply usually induces a change in the direction of their evolution (Sánchez-Arcilla and Jiménez, 1997), especially if other processes intrinsically linked to deltas such as subsidence are taking place. Even without considering the additional effect of RSLR, many deltas in the world have experienced such changes in their behaviour (which can be illustrated by a shift in the position of the river-wave-tide plane (Jiménez *et al.*, 1997a, 1997b & 1997c) for the Nile delta (Stanley and Warne, 1993) and the Ebro delta (Jiménez and S.-Arcilla, 1993)). As an illustration, the average annual input of sediments for the Ebro delta has been drastically reduced from about $30 \cdot 10^6$ metric tons per year to $0.1\text{-}0.2 \cdot 10^6$ metric tons per year (Palanques *et al.*, 1990). This reduction has been mainly associated with the construction of dams in its course which presently regulate 97% of the drainage basin (Varela *et al.*, 1986). The total sediment transport -bed and suspended loads- in the Po delta has experienced a less dramatic reduction from about $20.5 \cdot 10^6$ metric tons per year to about $7.8 \cdot 10^6$ metric tons per year, and much of this decrease has been associated with dredging and sediment extraction in the river course (Ruol, 1996).



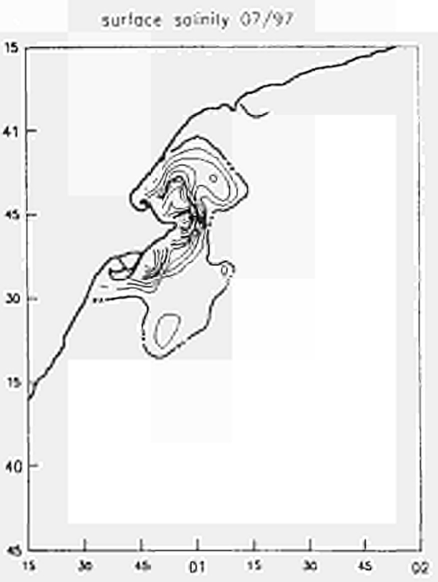


FIGURE 4.15 Illustration of the Ebro river plume features and their variations with seasonality, as observed in the MAST-III FANS project (Salat et al., 1998).

4.2.10 Estuary/river mouth processes

River mouth processes, integrating marine and riverine factors, constitute an important link in the overall coastal system (since e.g. all terrestrial sediment and nutrient inputs to the coast must go via the river mouth(s)). The relative balance between marine and riverine factors depends on climate change and obviously modifies river mouth processes. The energy level of river discharges and marine "mixing" factors determines whether the mouth behaves as a well mixed, partially mixed or salt-wedge estuary (see e.g. Dyer, 1986).

As an illustration, the Ebro river, under present conditions, features a Northern or principal mouth which behaves as a salt-wedge estuary for average discharges. This mouth coexists with a number of smaller mouths which however contribute a non-negligible amount of suspended sediments and nutrients (FANS-Bangor, 1998).

The change from river dominated to wave dominated deltaic systems, as observed in the Mediterranean, creates a substantial change in river mouth depositional processes. When the system is river-dominated, the deposits are basically a function of flow inertia (associated to the out-flowing discharge), bed friction and buoyancy of the fresh water mass. The significance of each factor depends on the out-flow velocity, the presence of density stratification in the receiving sea water and the outlet geometry, related to the width and depth of the receiving basin (Wright and Coleman, 1973). Based on this mechanism, out-flow models can be classified as full turbulent jet (dominated by inertia and frequent in relatively deep basins), friction dominated (plane) turbulent jet (dominated by bed friction and frequent in shallow basins) and a buoyant or hyperpycnal jet (corresponding to a receiving basin with strong vertical density gradients).

An increase in mean water level, due to an accelerated eustatic rise or to subsidence, will favour -for enhanced river discharges- inertia dominated river jets or -for reduced river flows and high mean water levels- buoyant discharges.

The conversion from river dominated to wave dominated deltas is directly associated with the relative increase in wave activity in front of the river mouth. As a result it can be anticipated that there will be a stronger deceleration and lateral expansion of the river discharge, an increase in the mixing processes, in particular related to wave breaking, and a more rapid deposition of sediment in front of the river mouth (see e.g. (Wright, 1985)). Waves, wave-driven flows and the associated "littoral drift" will be re-working the river mouth deposits much more rapidly. The (subaquaous) levees will thus be reworked into swash platforms and spit-like bar features. Likewise the change from a fluvial dominated to a tide dominated deltaic regime will be likely to produce a much stronger vertical and lateral mixing of the discharged fresh water mass (reducing the effects of buoyancy) and the development of more pronounced bi-directional sediment transport patterns at the river mouth.

4.3 COASTAL RESPONSES

4.3.1 Introduction

The significance of climatic change and, in particular, sea level rise is apparent in many domains but is particularly dramatic in the coastal zone where risks of flooding, coastal erosion, etc. are more intense (see e.g. (SCOR, 1991); (Koster and Hillen, 1995); (Stam and van der Weide, in press)). This statement is justified by the delicate dynamic equilibrium between geometry/distribution of depositional bodies in the coastal zone/shelf and the driving climatic factors (energetic input from waves, tides, currents and rivers) (see e.g. Nelson and Maldonado, 1990). Nevertheless, this equilibrium has been "at work" at a variety of scales.

The Holocene transgression has been an important control in the landward migration of the coastal zone, which determined the extent of the continental shelf. About 5,000 years ago most of the present coastal systems were initiated when the sea reached a level close to the present one. Because of this the near surface shelf deposits are transgressive and highstand system tracts, developed during the last sea level rise and the present conditions (see for example Bruun, 1988; Posamentier *et al.*, 1988a; 1988b).

As a consequence, present major deltas occur offshore from the principal sedimentary sources, while the coastal and shelf zones are wide and progradational (Ferran and Maldonado, 1990). In contrast, the coastal zone and shelf are in general relatively narrow and with a reduced sedimentary blanket where the sediment input is limited and there is active thermal or tectonic subsidence. Moreover the increasing influence of human impact experienced nowadays, in addition to the natural factors, is starting to alter the sediment supply to coasts (dam and river flow regulation policies, coastal development and mining of coastal sands, etc.). These factors together with the inherent difficulties in predicting sediment motion, transport and the associated morphodynamics, introduce uncertainties well in excess of those considered in the driving factor section. That is why, after presenting some ideas on the conceptual framework to approach "coastal responses", this section only deals in a summarized manner with coastal behaviour at two scales: "historic" and medium-term.

4.3.2 Conceptual approach

Coasts are geomorphological features resulting from the existence of a dynamic balance between the characteristics of the continental drainage basin and those of the marine receiving basin. The relative strength of these two systems determines the dominant processes governing the characteristic evolution of the coast. The analysis of coastal changes implies, as a consequence, a broad spectrum of processes, time and space frequencies, on environments which need to be considered as integrated systems (S.-Arcilla and Jiménez, 1997; Maldonado, 1997).

When analysing short and long term coastal processes, the main components of the coastal system are the drainage basin (including the characteristics and nature of the sedimentary material) and the coastal zone (including the characteristics of waves, currents and mean water level variations) (see for example SCOR, 1991; Bondesan *et al.*, 1995; Jiménez *et al.*, 1997a). Within this framework, the continental shelf must be considered as the seaward component of the coastal zone for the driving terms but also from the standpoint of evolutionary processes active in these two provinces (Maldonado, 1995).

At a longer time scale the evolution of coastal systems is more heavily influenced by additional factors such as tectonics, subsidence and sea level cycles which may have been the main controling factors of many coastal systems in the past at the long term scale (see for example Emery *et al.*, 1988; Nelson and Maldonado, 1990).

The application of the "littoral cell" concept for the study of integrated systems is particularly appropriate for short and long-term scales and for sedimentary deposits such as sandy beaches, where human influence is significant. A littoral cell is a natural sedimentological unit, such as e.g. the Nile delta-Levantine coast (see for example Golik, 1997) or the Ebro delta (see for example Jiménez *et al.*, 1997b).

These littoral cells are confined within an established "boundary condition", such as the presence of a mud belt which allows the coast to be considered "closed" for the coarse grain fraction of coastal sediments. The main controlling factor in the long term morphodynamic analysis of the cell is, then, the sedimentary balance. In this balance positive contributions are represented by terrestrial terrigenous input and cross-shore landward transport, while negative contributions comprise coastal sinks, mostly lagoons and dune fields, offshore sinks such as submarine canyons and relative sea level rise.

The mid term evolution of the coastal cell, in contrast, is largely controlled by shore processes driven by waves, currents and shorter term sea level oscillations, which determine sediment transport mechanisms and morphodynamic processes at those scales (see for example Jiménez and Sánchez-Arcilla, 1997).

This approach can also be described schematically in terms of sedimentary fluxes at different scales (FANS-Bangor, 1998), ranging from the high frequency variation of sediment fluxes associated with bed form dynamics in the nearshore and shoreface to the historical time scales associated with the overall coastal system fluxes. The four main components usually considered when analysing coastal responses are: 1) Geological-historical scale, 2) Long term scale, 3) Mid term scale and 4) Episodic events (Sánchez-Arcilla and Jiménez, 1997).

These different scales need to be identified and the coastal processes corresponding to each frequency band isolated. The main flux, from a morphodynamic standpoint, is the sedimentary flux, which controls the corresponding balance and is determined by the sediment supplies (amount, type and location) (see for example S.-Arcilla and Jiménez, 1997; Maldonado, 1995). Traditionally the rivers with a major drainage area have been considered the main sedimentary source for the coastal environment. Although alternative sources such as the erosion of cliffs and coasts and the shoreward erosion/regression of shelves are becoming recognised as more and more significant.

Human influence is apparent when looking at the sedimentary supply from rivers, for example caused by deforestation and overgrazing of river drainage basins in the Mediterranean, during the late Holocene. Damming along many rivers has resulted in a significant loss of sediment supply to the continental margin during the past few decades. These effects are probably opposite since increasing deforestation and land erosion tends to enhance sedimentary input to the coast, while damming and river flow regulation play the opposite role (Figure 4.16). For instance the Nile delta after construction of the Aswan high dam is suffering severe coastal and deltaic erosion, similar to that now beginning to be observed in the Ebro delta or in many other Mediterranean deltas (Stanley and Warne, 1993; Jiménez *et al.*, 1997a & 1997c; Poulos and Chronis, 1997). This is an increasing and severe problem for the entire Mediterranean and similar seas, since most of the sediment input has ceased (Nelson and Maldonado, 1990). As a result deltaic erosion, salt water intrusion and cropland destruction are increasing, as shown in Figure 4.16 The effect of sediment contamination on the riverine sedimentary supplies and continental margin depocenters is starting to be well documented and will also be sensitive to any change in climatic conditions.

The availability of sediment thus controls the progradation of coastal systems (well linked to the periods with greater supply of coarse sediments). The location and nature of sediment supplies are also important controls, quite sensitive to climatic changes. As an illustration, linear or point sources determine growth patterns and depositional rates, since point sources generally tend to generate coastal deltas and estuaries while linear sources normally result in sandy barrier beaches.

On the other hand, sea level changes influence the nature of sedimentary supplies and, in consequence, the resulting evolution of coastal systems (apart from the effect that sea level has on driving factors as presented in the previous section). During the Holocene transgression sedimentation rates were low as a result of the landward displacement of the coastal systems (Nelson and Maldonado, 1990). In contrast, when the present sea level was reached, the coastal systems prograded offshore and sedimentation rates increased, particularly off the sedimentary point sources where deltas began to grow (see Figure 4.14, above). The coastal response, in general, has been site-dependent, controlled to a significant extent by local factors.

The sedimentary sinks intersecting coastal cells are also a significant factor since they act as traps for marine materials extracted from the sediment budget. Important sinks include outer shelf canyons, coastal lagoons and the inner shelf below wave action. This loss in sediment availability has led, in the late Holocene, to serious erosion problems, which combined with the expected acceleration in climate change may be particularly serious for low lying coastal areas.

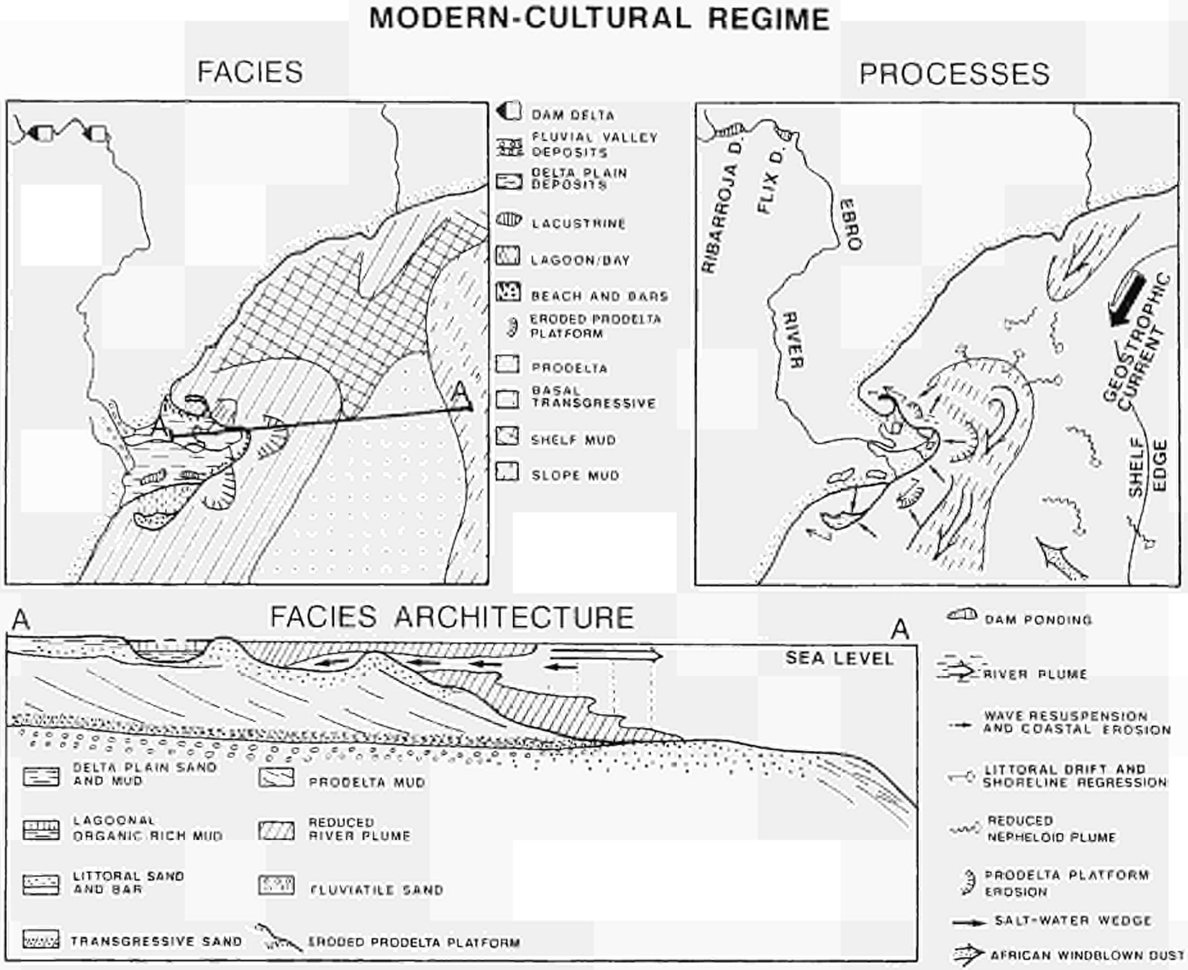


FIGURE 4.16 Generalised sedimentary facies and processes of the present human-cultural regime in the Ebro coast and shelf, showing the significant influence of human impact due to reduced sediment supply after dam construction (from Nelson and Maldonado, 1990).

4.3.3 Geological/historical responses

The relative importance of the factors controlling coastal changes depend on the scale of the analysis of a particular coastal cell. These scales have a certain degree of subjectivity, given the present level of knowledge, which results in a relatively large diversity of proposed classifications (see for example de Vriend, 1991a, 1991b; Fenster *et al.*, 1993; Crowell *et al.*, 1993; Jiménez *et al.*, 1997a & 1997b). Historical processes are controlled by the characteristics of the drainage and receiving basins (see for example Nelson and Maldonado, 1990; Crowell *et al.*, 1991). The historical/geological processes considered in this section are associated with typical scales of hundreds to thousands of years and can be illustrated by sea level and climatic cycles, tectonics and the characteristics of the terriginous input. Long term processes have been associated with changes at a time scale of decades and the spatial scale of the littoral cell or the complete coastal system (see e.g. (Jiménez *et al.*, 1997a & 1997c)). The main changes at this scale occur in the morphology of the coast and in the sedimentary budget, being longshore sedimentary transport processes one of the main factors for coastal reshaping.

When considering the coastal response to climatic variations at the geological/historical scale (the longest one), the main controls are those related to changes in continental sediment supplies and those in the receiving basin. The first, as previously stated are verified in the drainage basin and can be due to a change in the river liquid discharge and/or in the soil cover over the basin. Generally speaking these changes are usually linked although they produce opposite effects.

As an illustration, a decrease in rainfall regime would induce a decrease in the liquid river discharge which leads to a decrease in the sediment load able to be supplied by the river to the coast. Since sediment is the limiting factor for coastal stability, this should result, under a steady wave and current climate, in an increase of reshaping and reduction processes.

On the other hand, a rainfall decrease would be accompanied by a decrease in the vegetation cover. This naturally-induced deforestation process should result in a larger potential erodibility of the soil and, thus, in a larger potential sediment availability to be supplied to the coastal system. As both induced effects are opposite, which of them would dominate will depend on the overall effects of rainfall decrease. Different effects would then be dominant, depending on the change of average extreme rainfall events.

However these effects are not likely to be produced, at least in the positive direction, i.e. to promote a larger sediment availability. This is due to the fact that, since these changes are verified in the drainage basin, the only way to reach the coast is by using the river as a vector. Since presently most of world-wide rivers are largely regulated by reservoirs and barrages, any increase in sediment supply will be counteracted by their blocking effects. On the other hand, the increasing water exploitation for human activities is inducing a decrease in circulating liquid discharges, at a much shorter term than that associated with climate change. Because of this, it is expected that human activities in drainage basins should be much more effective than climate change in affecting coastal processes.

The other main climate control in coastal response, at this scale, is the change in relative sea level rise. Sea level change effects on coastal systems (at this scale) are one of the most cited topics examined in recent literature.

Most of the existing approaches are based on an equilibrium assumption, namely that the coastal system is at equilibrium under the present marine climate and that any change in sea level would induce a vertical and horizontal shift of the system, tending to maintain that equilibrium shape. If there is a sea level rise, the coastal response is transgressive with shoreline retreat and a profile shifting upward (see for example Bruun, 1988; S.-Arcilla *et al.*, 1995; Stive and De Vriend, 1995; Jiménez *et al.*, 1997b & 1997c). On the other hand, a sea level drop would result in a regressive response with shoreline advance. The kind of response depends not only on the direction of the change but also on its celerity.

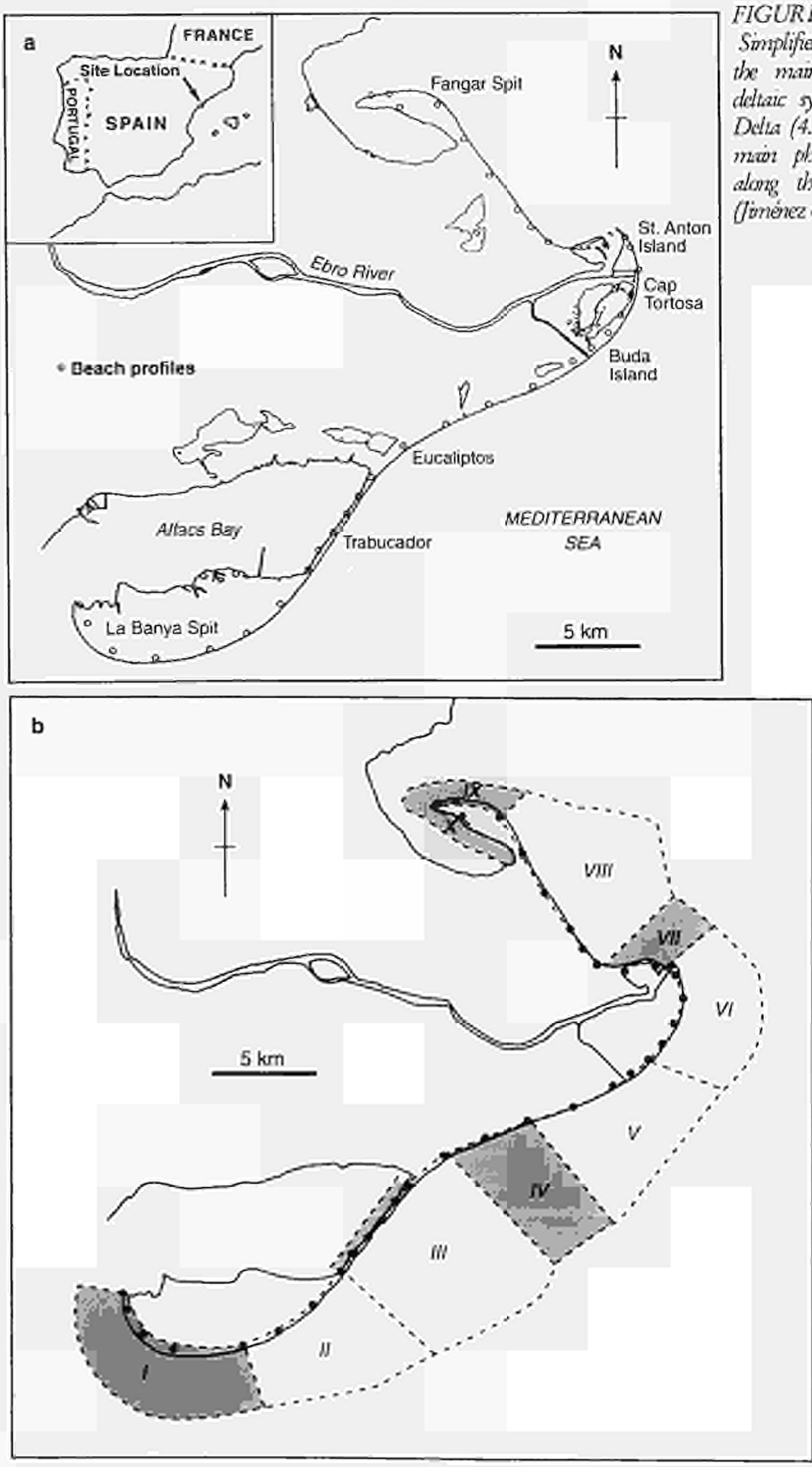


FIGURE 4.17
Simplified image of one of the main Mediterranean deltaic systems, the Ebro Delta (4.17a) showing the main physiographic units along the coast (4.17b) (Jiménez et al., 1996).

As a consequence, for a gradual and slow sea level rise, if the coastal system is still fed by riverine supplies it would be able to cope with the change. The paradigm of such adaptation response is a deltaic system (Figure 4.17). If sediment is being supplied to the deltaic plain, the vertical accretion will counteract the change in relative elevation, whereas sediment supplied to the coast will be used to maintain the coast in place. This stresses the links between the different changes, because they can act in a synergetic manner. This means, for instance, that sediment supply has to be also considered when sea level changes effects are studied because, at the end, coastal stability is the result of the sediment budget.

These changes in sea level can be also accelerated due to human activities such as underground exploitation. Such induced subsidence may significantly increase sea level rise (e.g. Sestini, 1992).

4.3.4 Medium/short-term responses

i) Due to free-surface ocean waves

The accurate quantification of wave-induced sedimentary transport and morphodynamics is difficult, or even impossible, in spite of recent technological advances. Sediment transport, subdivided into bed and suspended load transport, can be assessed with different degrees of certainty depending on the time/space scale and the type of coastal environment. Suspended load transport can be "measured" with mechanical, optical and acoustic devices in combination with current meters for limited periods of time and limited spatial coverages. Bed load transport is still difficult to quantify from direct observations, both in nature and in the laboratory.

The three dimensional character of wave-induced morphodynamics is also difficult to assess and quantify, which has resulted in the frequent simplification into 2DH or 2DV, "pictures". This results in a decoupling between longshore and cross-shore processes, even though the relation between horizontal and vertical circulation/transport patterns is only partly understood (see for example Wallace, 1993).

Because of the non-linearity of sediment transport processes -in terms of the driving hydrodynamics-, the wave-induced coastal response is dominated by the more energetic events with the lowest frequency of occurrence (although this statement is still not universally accepted). For the purely wave induced coastal response this question can only be solved once sufficiently long and detailed observations become available. An example of such an approach for sediment transport processes (for wave induced sediment transport processes in the nearshore) appears in Figure 4.18. This example is based on wave induced sediment transport measurements and computations along the barrier island coast of Terschelling (Houwman and Ruessink, 1997; Ruessink *et al.*, 1998).

The sediment transport rates, computed by using long term flow measurements, are related to the local relative wave height (H_{m0}/h) at that specific location in order to take into account the variations of water level (tide and wave induced) and the local wave height. The various transport components (bed load, mean and oscillatory suspended load transport) show a clear dependence on the relative wave height. Since water levels and wave height have been measured over periods of more than one year, the probability density function of the relative wave height (H_{m0}/h) can also be determined and plotted in the diagram. The relative importance of transport processes in relation to boundary conditions can be evaluated by multiplying the transport rates with the associated values of the probability density function. Maximum transport rates for an average year appear to be related to intermediate values for the relative wave height (of about $H_{m0}/h=0.35\text{-}0.45$) and do not occur for the maximum observed relative wave height.

This same approach could be used for other types of coasts, although the relationships will be different. In most cases, however, long term records of parameterised boundary conditions are not

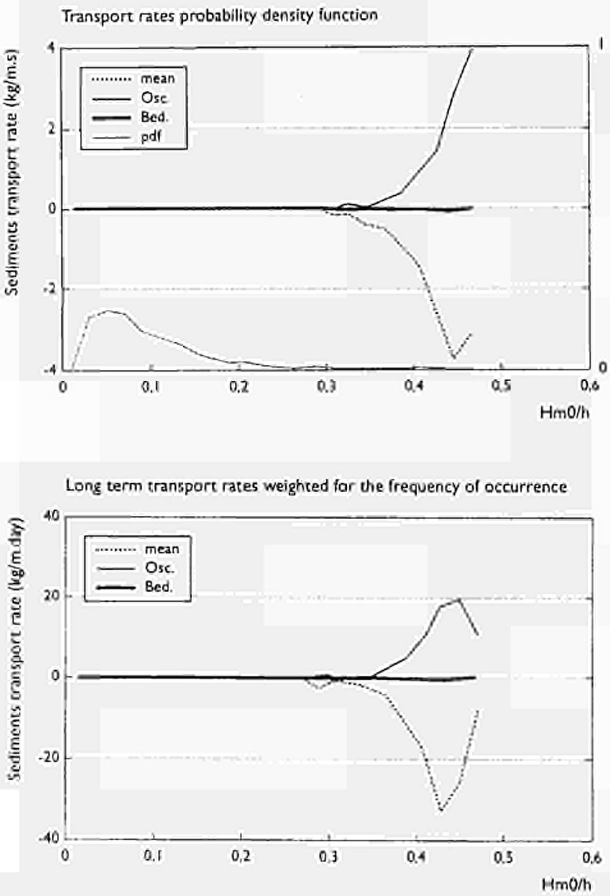


FIGURE 4.18 Sediment transport rates (bed-load and suspended oscillatory and mean load components) as a function of the wave height H_{m0}, together with the H_{m0} probability density function (18a). The long-term transport rates (weighted by the frequency of occurrence) shows the relative scanner/ bekenskamor contribution of each wave class (18b) (Houwman and Ruessink, 1997).

available and/or the links between the detailed morphodynamic processes and these boundary conditions are not well established. The same applies to the existing morphodynamic models, where the non-linearity of the governing equations (illustrated e.g. by the fact that the combination of waves and currents generates composite bed forms and creates a bed roughness markedly different from bed forms and roughness due to either waves or currents alone) make the modelling task more difficult, sensitive to inaccuracies and uncertainties and even limits the predictability (see e.g. Capobianco *et al.*, 1996a; De Vriend, 1991a; 1991b).

The present generation of process-based morphodynamic models still suffers from a "knowledge gap" in relation to several important constituent processes, which have not yet been implemented (a typical example are swash zone processes, important for longshore and profile evolution models, but which are not normally considered). Many aspects of the models' validation still remain open to question, since there are so far no data with the adequate spatial and temporal resolution and the laboratory data show important limitations for a proper validation. Most of these process-based models incorporate modules which still require a substantial number of empirical data sets (wave-induced sediment transport formulations, with efficiency parameters for the various load components illustrate this point).

The interaction between morphodynamics and hydrodynamics at this medium-term scale also remains a somewhat open question, because of the uncertainties in driving factors, the limited knowledge on sediment transport processes and the corresponding interactions (see for example de Vriend, 1993). That explains why this type of process-based morphodynamic model cannot be used to predict coastal behaviour at the longer time scales so as to assess the effects and interactions with climate (in particular sea level) changes. The use of the present generation of process-based models for decadal scales is therefore not possible since the accumulated errors may well exceed the order of magnitude of the predicted coastal response.

Within this framework it is easy to understand the difficulty of quantitatively assessing the effect of climatic change. Even a minor change in the average or dominant wave direction will alter the magnitude and direction of the associated longshore sediment transport flux, and will initiate an adaptation of the shoreline orientation, as well as a shift of areas of sediment convergence and divergence (convergence corresponds to depositional areas and divergence corresponds to erosional areas). This can be illustrated by the drift currents along the wave dominated coast of the Ebro and Rhône deltas in response to different wave regimes (Suanez and Provansal, 1996; Jiménez and Sánchez-Arcilla, 1993). The present Rhône delta system, for instance, is characterised by a limited input of fluvial sediments and consequently a greater dominance of the wave induced drift system. The same, as mentioned before, applies to the Ebro delta which results in an enhanced sensitivity to any change in the wave driving factor.

ii) Due to wind events

Although many of the ideas just presented are also applicable to wind-induced processes it is worthwhile to make some specific remarks.

Changes in wind velocity and direction will affect aeolian processes and the corresponding dynamics in different ways. The main controlling parameters are the sediment grain size, the geometrical properties of the coastal domain (beach length, width, slope, etc.) and the wind regime. Wide dissipative beaches favour aeolian processes and, as a result, various authors have suggested a direct relationship between shoreline morphodynamics, aeolian processes and dune formation (see e.g. Short and Hesp, 1982). This implies that changes in the wave climate, which result in changes in the beach state and geometry (although indirectly), will also affect the result of aeolian processes.

In a more direct manner, the wind regime is important for the entrainment of sediment and the actual transport rates and direction. An increase (decrease) in wind velocity may lead to an increase (decrease) in sediment transport rates, with the development of blow- outs and the landward migration of coastal dune fields (see e.g. Hesp and Thom, 1990). A change in wind direction may, thus, result in a somewhat different effective "fetch" of the beach, which will subtly affect aeolian processes. In some coastal areas dune formation is the product of alternating onshore and offshore winds. In the Rhône delta, for example, the wind climate is determined by the predominant Mistral, blowing from the NW (offshore), and the Marin, blowing from the SE (Termaat, 1996). A change in the balance between these two opposing winds may seriously alter the aeolian sedimentary budget. The same applies to the Ebro delta, where the dominant winds are from the NW and NE (see for example García *et al.*, 1984).

In many cases the relationships between climate induced changes in the wind field and coastal processes may be much less direct than the ones suggested in here. A change in wind regime may be part of a more regional climate change which is also reflected by changes in rainfall and vegetation. Since soil moisture is a limiting factor for aeolian transport (see for example Jiménez *et* *al.*, 1999) and vegetation is also a stabilisation factor for dunes which will limit dune movement (see for example Jiménez *et al.*, 1999) climate changes altering these variables would influence aeolian induced processes. A clear example of the influence of the regional climate on aeolian transport and resulting dune evolution has been recently presented (Jiménez *et al.*, 1999). These authors have analysed dune behaviour in NE Brazil and they have found a pulsating dune movement pattern as a function of the regional seasonality in rainfall and wind intensity. During the wet season dunes are temporarily at rest, due to the decrease in wind intensity and the prevailing high rainfall, whereas during the dry season, the absence of rainfall and the increase in wind intensity drive a free dune migration (Jiménez *et al.*, 1999). Any change in regional climate would lead to a change in yearly effective aeolian transport and dune migration.

4.4 CONCLUSIONS

For many coastal systems the "human factor" is presently a far more important condition/control than a gradual or even accelerated -with the presently expected rates of acceleration- climate change or sea level rise. As an illustration, in the Ebro delta, and within the framework of the MEDDELT Project (Environment Programme-Commission of the European Communities), it was found that the management policies will dominate over climate changes with respect to geomorphic "formation" processes in the whole deltaic area (Sánchez-Arcilla *et al.*, 1996b).

The delicate equilibrium which determines coastal morphodynamics makes low-lying coastal areas an ideal test case to study the disturbance produced by human or climate forcing (or a combination of the two which are, in any case, difficult to separate in reality).

Sedimentary budget and fluxes in the various coastal domains (and in particular in the shelf, nearshore and surf domains), are essential to understand the long term behaviour of a coastal stretch and to allow a better management compatible with human interference. The extraction of aggregated indexes (e.g. sensitivity or susceptibility indexes such as those proposed in Sánchez-Arcilla *et al.*, 1998) can be a means to filter out part of the present uncertainties in drivings and responses, and also to breach the gap between the processes-based research on coastal dynamics and the practical applications in the hands of coastal zone managers.

Even for process-based models (conceptual or predictive), their validity in scenarios with very limited sedimentary supplies and an accelerated rise of sea level is doubtful, given the amount of empirical information built in the present generation of models.

The non linearities present in the coastal system (both in individual components such as sediment transport formulations and also in the aggregated behaviour system as illustrated by the hydrodynamic-morphodynamic interactions) make it desirable to have large enough data bases, adequately covering the most energetic (stormy) events, which are believed to be the main controlling factors from a morphodynamic standpoint. This does not preclude coastal response varying significantly from site to site depending on "local factors" (sediment availability, etc.).

The establishment or continuation of adequate field data sets is, thus, considered a "must" to advance the knowledge on climate change implications for coastal processes. The expected variations in drivers and responses are very small at the yearly scale (the expected eustatic rise in sea level, considering the most pessimistic estimates, is in the order of millimetres per year). This means that the accuracy of instruments and the quality control of field campaigns should be carefully assessed from the beginning since, otherwise, the observational efforts may be totally wasted. The observational policy and campaign strategies will also vary depending on whether it is a process-oriented field campaign or a long-term "averaging" one. On the other hand the maintenance of this prolonged field effort should be emphasised since the short term benefits may not be clearly apparent for many of the interested parties.

Acknowledgements

The authors would like to acknowledge the following persons, with institutions and research projects, who have in one way or another contributed to this text:

Mr. Graham Alcock (Proudman Oceanographic Laboratory, U.K.)
Dr. Cesar Andrade (Faculdade de Ciencias, Universidade de Lisboa, Portugal)
Prof. Giuseppe Bendoricchio (Università di Padova, Italy)
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5
IMPLICATIONS FOR COASTAL ZONE MANAGEMENT

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Coastal Zone Management (CZM) is not a new concept. It is a methodology that came into legislative existence as a result of increasing human pressure on the usage and living space of the barrier islands off the USA eastern seaboard. The explicit recognition of a need for management of the coastal zone is shown in the USA's Coastal Zone Act (1972), the first overt national CZM legislation. The recognition of contemporary allogenic forcing (i.e. the accelerated greenhouse effect inducing changes in both relative sea-levels and storminess patterns, frequency and magnitude around the world) and its effects on the coastal zones of the world has increased the need for sustainable CZM in the next millennium. It has become increasingly clear that relative sea-level changes will result not only in a variable response at the coast, depending on its typology and geomorphological characteristics, but also will have a profound impact on the physical and cultural landscape - hence the need for CZM. Given that future allogenic change will lead to coastal change, the management of such change will become the central remit of coastal zone management of the next millennium.

This chapter intends to analyse in a CZM context, the impacts and hazards to the coastal zone in terms of coastal sensitivity (section 5.1), and the possible responses to be found to these threats (section 5.2). In two further sections, the position of CZM in the modern physical and cultural context is analysed (section 5.3), followed by an analysis of future research needs in Europe, in order to work towards a successful implementation of (integrated) CZM (section 5.4).

5.1 IMPACTS AND HAZARDS

5.1.1 Sensitivity: A definition

Sensitivity refers to the propensity of a system to respond to a minor external change. The change occurs at a threshold, which when exceeded produces a significant adjustment. If the system is sensitive and near a threshold it will respond to an external influence, but if it is not sensitive it may not respond (Schumm 1991).

All dynamic systems in the world are subject to changes and possess their own, specific, degree of sensitivity to external forcing. The climate system's long-term sensitivity responds to the Milankovic cyclic variations. On the short-term it responds to variations of amounts of greenhouse gases in the atmosphere. The ecosystem's sensitivity responds amongst others, to climate shifts, and the morphodynamic system's sensitivity responds amongst others, to sea-level changes, tectonic activity, or climate forcing.

As can be expected from the definition above, and as each system is in dynamic (metastable) equilibrium, there exists a bewildering array of forcing factors on the one hand, and levels of thresholds on the other hand. In fact, dynamic metastable equilibrium is mostly controlled by negative feedback mechanisms that ensure as much as possible a dynamic *status quo* of the environment until external forcing becomes so great as to push the system's resilience beyond the equilibrium threshold. Positive feedback mechanisms then force the system into a new dynamic equilibrium, generally accompanied by, e.g. significant ecological and/or morphodynamic changes (Figure 5.1). In natural systems, such changes are gradual due to the complexity of internal forcing/response mechanisms that control the thresholds, and to the lag time necessary for the system to adjust to the new dynamic equilibrium. External events however, may accelerate or overwhelm the internal mechanisms of adjustment. Natural (sudden) forcing like volcanic eruptions, earthquakes or major storms, and human forcing like deforestation or the exploitation of sensitive coastal zones, may result in a rapid crossing of equilibrium thresholds, and the lagged response of the system is replaced by an immediate response.

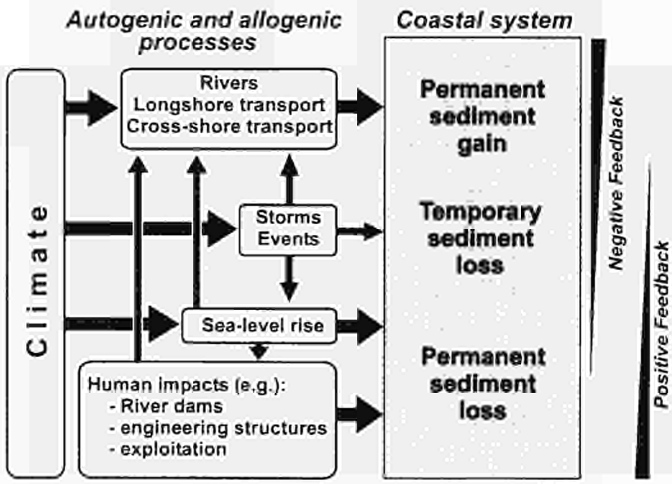

FIGURE 5.1 Example of forcing processes on the sediment budget to and from the coastal system. Threshold boundaries determine where and how feedback occurs.

As an example of the first type of sudden forcing in a coastal zone setting, the 1694 Culbin Sands disaster in north-east Scotland can be mentioned (Lamb 1991) where - as a result of one single violent storm - 20 to 30 km^2 of farmland where overwhelmed by moving dune sand. The area remained a desert of wandering sand for 230 years until its successful afforestation from the 1920's onwards. This same event may serve also as an example of the second type of forcing, as it is surmised that exploitation of the marram grass growing on the sand dunes for roof thatching, left the sand hills more vulnerable to storms.

From this example one can derive also a word of caution about the interpretation and/or extrapolation of sudden threshold crossings, especially when dealing with reconstructions from the past. One should be fully aware that in sensitive cases, minor inputs might cause major changes, which elsewhere may have little or no effect. Such local changes, if they left a stratigraphic record, could easily be interpreted as the result of a major climatic or morphodynamic change, which would be a totally incorrect assumption (Schumm 1991).

Thus, space and time scales have to be considered as major constraint factors in interpreting coastal process forcing (Cowell and Thom 1994; de Groot 1996).

5.1.2 Major threats to heightened sensitivity

If sensitivity can already be under stress from minor changes in the geosystem, how is sensitivity going to respond to major changes? Both increases in relative sea-level rise and changes in storminess patterns and intensities resulting from the accelerated greenhouse effect (Titus 1987), set the stage for major morphodynamic changes in the coastal systems of the world. The awareness of these changes has become widespread through the IPCC reports (Houghton *et al.* 1990, 1992 & 1996). As a consequence, a large number of research projects were instigated, many of them under the auspices of the Commission of the European Communities (e.g. EPOCH, Environment, MAST). Further work has been generated by international groups (e.g. Tooley and Jelgersma 1992; IGBP-LOICZ: Pernetta and Milliman 1995), or national ones (e.g. the Coastal Genesis Project in the Netherlands: Beets *et al.* 1992, 1994; Van Rijn 1995). It is impossible to give an overview of all these projects within the scope of this chapter, however, summaries of such work can be found elsewhere (e.g. Nolan 1995; Ter Mors 1996; Weydert 1996).

The balance between susceptibility (i.e. the impacts of a large array of different forcing factors) and resilience (i.e. the propensity of the system to adjust to these forcing factors) identifies the sensitivity of the system. Highly sensitive systems are prone to reduced or reducing resilience and are more likely to be associated with energy thresholds that are crossed, and hence allow major changes in the coastal zone, with all the associated consequences for coastal ecosystems and human communities. Concern about coastal resilience is heightened, in the light of possible associated rapid self-organisation in coastal systems at times of threshold change, that forces major disruptive changes to spatially mature and established coastlines (e.g. Orford *et al.*, 1996).

If relative sea-level rise and changes in the Earth's storminess patterns are major threats to the world's coastal system, then human pressure on most coastal areas is another major threat to coastal sensitivity. This situation has mainly been dictated by the cultural/economic encroachment from the landward side of the coastal zone, first by the historic and contemporary economic development of coastal areas, secondly by the growing tourism and recreation pressure since the latter part of the 19^{th} Century (see section 5.3). As a result of this human pressure, the stimulus for coastal management has been focused on protective responses to coastal change that tend to heighten sensitivity, to the detriment of more proactive and adaptive approaches that reduce coastal sensitivity.

All this being said, it is the purpose of this section to assess the balance between susceptibility and resilience of coastal systems as a function of its sensitivity in the light of major impacts, and the possible responses that can be generated.

5.1.3 Controlling factors

From the exposition above of mechanisms controlling natural coastal systems, it is necessary - for a better understanding of the morphodynamic processes that control coastal systems - to assess what are the feedback mechanisms, the sensitivity levels, and the thresholds that can be recognised.

Coastal systems are complex and subject to a vast array of forcing mechanisms that can be subdivided into five main groups (Figure 5.2). All groups are linked to each other by intricate

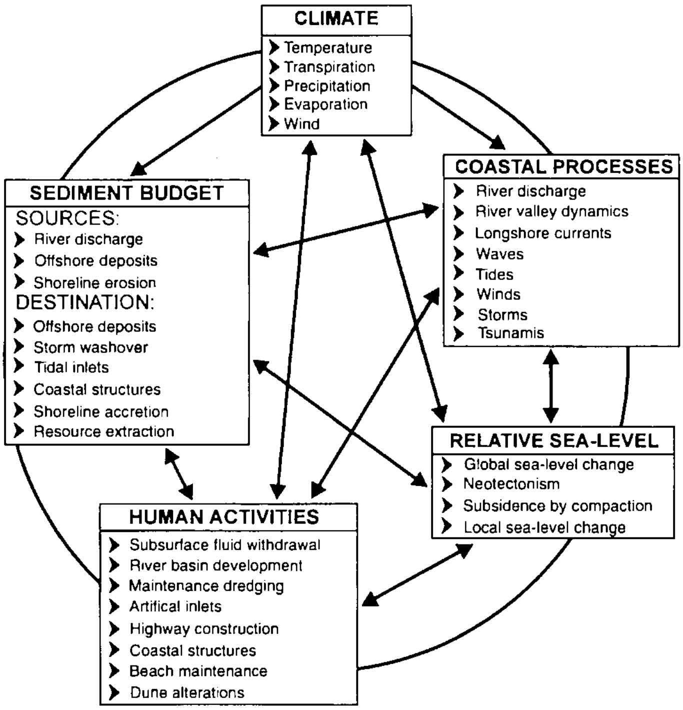

FIGURE 5.2 Interaction of different forcing factors on coastal systems (after Williams et al. 1991).

feedback mechanisms. In Table 5.1, an assessment is made for each group in terms of sensitivity level (high to low), most probable feedback (negative or positive), and threshold (low to high), directly in relation to the coastal systems. It should be noted however, that many if not all of the assessments are dependent on forcing from other groups so that a low sensitivity, e.g. in coastal processes, may turn into a high sensitivity by the forcing from human activities. None of the factors mentioned are standalone elements within particular coastal zones. Also, the type of coastal system may result in different values of sensitivity, feedback or thresholds!

Table 5.1 mainly reflects the situation on predominantly sandy coasts and deltas. Tentatively, it can be surmised that gravel beaches and cliffs will have slightly lower sensitivity values (the lower extreme being rocky coasts), and that tidal flats and saltmarshes will show slightly higher sensitivity values.

This being said, an assessment can be made about which environmental parameters have the highest impact on coastal systems and, in the case of climate change, what factors or combination of factors are most prone to impact on coastal systems. Finally, an evaluation can be given about the degree of impact on coastal systems' sensitivity and whether climate (driving sea-level changes amongst others) or other factors are dominant.

	Sensitivity	**Feedback**	**Threshold**
Waves (cross-shore currents)	medium/high	negative	medium/low
Longshore currents	medium	negative	medium
Tides	low/medium	negative	medium
Rivers	low	negative	high
Winds	medium	negative	medium
Storms	high	negative/positive	low
"Events" (e.g. tsunamis)	high	positive	low
Sediment Source	**Sensitivity**	**Feedback**	**Threshold**
Rivers	medium/high	negative	medium
Shoreface/Shelf	high	negative	low
Shoreline erosion	high	negative/positive	low
Sediment Sinks	**Sensitivity**	**Feedback**	**Threshold**
Cross-shore accretion	high	negative	low
Storm washover	high	negative	low
Tidal inlets	medium	negative/positive	medium
Shelf	low	positive	high
Relative Sea Level	**Sensitivity**	**Feedback**	**Threshold**
Tectonic subsidence	low/medium*	negative/positive	medium
Tectonic uplift	low/medium*	negative/positive	medium
Compactional subsidence	medium	negative	medium
Global sea-level change	medium/high*	negative/positive	low/medium
Local sea-level change	high*	positive	low
* *depends on rates of change*			
Human activities	**Sensitivity**	**Feedback**	**Threshold**
Coastal structures	high	positive	low
Resource extraction	high	positive	low
River basin development	medium/high	negative/positive	low/medium
Channel dredging	high	positive	low
Beach maintenance	high	negative	medium
Dune alteration	high	positive	low
Tourism	medium/high	positive	low/medium
Subsurface fluid extraction	high	positive	low
Climate	**Sensitivity**	**Feedback**	**Threshold**
Temperature	low/medium	negative	high
Transpiration	low	negative	high
Precipitation	low/medium	negative	high
Evaporation	low	negative	high
Variation of storm intensity	medium/high	negative/positive	low/medium
Shifting of storm belts	medium/high	negative/positive	low/medium

TABLE 5.1 Assessment of sensitivity, feedback, and thresholds upon different forcing factors on the coastal zone.

Particular cases of impact on all coastal systems are sudden events like tsunamis (see chapter 3, above). Although they are relatively rare and often concentrated in earthquake prone areas of the world, on a European scale a number of destructive tsunamis have been recorded in the past of which the tsunami associated with the Lisbon earthquake of 1755 is the best known. Studies on tsunami impacts show that all coastal systems may suffer more or less permanent damage from their occurrence (e.g. Atwater and Moore 1992; Bondevik *et al.*

1997; Dawson *et al.* 1988; Hindson *et al.* 1999; Long *et al.* 1989; Ribeiro 1994; Shennan *et al.* 1996; Shi *et al.* 1995; Young and Bryant 1992). Although there are no protection measures readily available, tsunamis are of such potential risk that their effects should be considered in coastal zone management assessments.

5.1.4 Coastal systems

Coastal systems and environments have been subdivided into a number of types based on their morphology and facies (e.g. Leeder 1982; Walker 1984; Carter 1988; Paskoff 1994). In the following sections, each coastal environment will be treated according to its sensitivity to climate change, or other environmental impacts in so far as this is relevant to the system in question. An overview of sensitivity impacts and possible responses is given in Table 5.2.

5.1.5 Impacts

Beaches, barriers and dunes

Sandy beach environments and barrier systems, including inlets and backbarrier areas, are especially susceptible to changes under external forcing. However, and with the exception of the post glacial sea-level rise, climate is not primarily a forcing factor of importance as long as sediment sources of enough volume (foreshore, river input, dunes) are available to replenish sediment losses through climate induced sea-level changes. Cross-shore transport by wave action plays here a dominant role in maintaining barriers.

However, the postulated thermal effect on the major ice caps of the world and the effects of oceanic thermal expansion on sea-level rise, form a possible hazard to be taken into account at the present level of knowledge. Although accelerated sea-level rise will result at first in an increased amount of available sediment (Jennings *et al.*, 1998), in the end a decrease of sediment supply towards the coast and especially within the barrier/backbarrier systems is expected (e.g. Oost and De Boer 1994; Donselaar 1996). In terms of preservation potential, the threshold towards barrier decay will be crossed when the amount of sediment supply required to maintain barriers exceeds the effective supply (Donselaar 1996). This is directly related to the rate of sea-level rise. Barrier decay will result in increased sedimentation in the backbarrier areas. However, since sedimentation commonly lags behind sea-level rise, a certain percentage of intertidal flats will change into subtidal areas (Oost and De Boer 1994), see also the effects of human influence below.

One other climate factor that potentially can be of direct concern to sandy coast sensitivity is the variation of storm intensity and/or the latitudinal shifting of storm belts. But again, this is (partly) compensated by a potential and sufficient sediment supply to the coast. Special attention however, should be paid to dunes as they may respond dramatically to changes in storm intensities and to changes in precipitation in particular (e.g. Heidinga 1984; Van Huis 1989; Doody 1991; Jelgersma, *et al.* 1995).

Human influence is, in many cases, a major forcing factor in these systems in a, paradoxically, destructive way (e.g. Granja *et al.* 1993). Especially in those areas with a low sediment supply, coastal protection measures can have a positive feedback effect on the evolution of sandy beaches and dunes (e.g. Williams *et al.* 1991; Granja 1999), with rapid threshold crossings towards permanent coastal erosion. Notorious measures are seawalls, revetments, breakwaters, and groynes (e.g. Ehlers and Kunz 1993; Paskoff 1994; Granja and Carvalho 1995) and, to a lesser degree, the fixation of sand in the foredune. However, erosion

	Sensitivity to climate change	Possible response	Sensitivity to other forcing factors (partly related to climate)	Possible response
Beaches and barriers	• Temperature: low/medium • Transpiration: low • Precipitation: low • Evaporation: low • Storm intensity variation: medium/high • Storm belts shifting: medium/high	• unknown • irrelevant • irrelevant • irrelevant • unknown • unknown	• sediment supply • sea-level rise • human influence (engineering)	• beach/foreshore nourishment • *idem* • dismantling of hard structures
Dunes	• Temperature: low/medium • Transpiration: medium/high • Precipitation: medium/high • Evaporation: medium/high • Storm intensity variation: medium/high • Storm belts shifting: medium/high	• unknown • plant cover stabilisation/protection; groundwater table control • *idem* • *idem* • unknown • unknown	• sediment supply • sea-level rise • human influence	• indirectly: beach/foreshore nourishment; directly: restrict dune fixation • *idem* • control/restrict access; education
Tidal flats, tidal marshes and lagoons	• Temperature: medium/high • Transpiration: low • Precipitation: low/medium • Evaporation: low/medium • Storm intensity variation: medium/high • Storm belts shifting: medium/high	• unknown • irrelevant • irrelevant • plant cover stabilisation/protection • unknown • unknown	• sediment supply • sea-level rise • human influence	• unknown • unknown • environmental protection measures; retreat of human occupation
Estuaries and deltas	• Temperature: medium/high • Transpiration: low • Precipitation: medium/high • Evaporation: low/medium • Storm intensity variation: medium/high • Storm belts shifting: medium/high	• see sediment supply • irrelevant • see sediment supply • irrelevant • unknown • unknown	• sediment supply • sediment bypass • sea-level rise • subsidence/compaction • human influence	• avoid damming of rivers; artificial bypassing to estuaries/deltas • unknown; creation of catchment areas • artificial protection of levees • artificial protection of levees • avoid damming of rivers, dredging and regulation of channels; restriction of fluid extractions and drainage
Cliffs	• Temperature: low • Transpiration: low • Precipitation: low/medium • Evaporation: low • Storm intensity variation: medium/high • Storm belts shifting: medium/high	• unknown • irrelevant • plant cover stabilisation/protection • irrelevant • engineering protection/nourishment of cliff base • engineering protection/nourishment of cliff base	• sea-level rise • human influence	• engineering protection/nourishment of cliff base • retreat from cliff tops

TABLE 5.2 Assessment of coastal sensitivity to climate change and other - partly climate related - forcing factors, and the possible responses.

mitigation techniques that closely replicate natural processes (e.g. beach nourishment, sand dune creation and protection, and shoreline re-creation) may provide temporary protection of endangered beaches and dunes (Van der Maarel 1979; Williams *et al.* 1991). However, only with the necessary background knowledge about natural processes acting on the beaches can these measures be successful (Pilkey *et al.* 1994). A particularly serious effect of human influence can be expected in the backbarrier areas as artificially fixed mainland coasts and/or the barriers themselves are anticipated to lead to sediment deficiency in the backbarrier region, and thus to a permanent drowning of substantial parts of the intertidal zone (Van der Spek and Beets 1992; Flemming and Davis 1994; Oost and De Boer 1994).

Tidal flats, tidal marshes and lagoons
Tidal flats, tidal marshes and lagoons have a particular sensitivity towards external forcing such as sea-level rise or storms. Already in the preceding section, the impact of an accelerated sea-level rise on backbarrier areas has been outlined. The same can be said of those flats and marshes occurring in other areas like embayments or estuaries, without the relative protection of a barrier. Marsh development has been shown to be controlled by changes in tidal amplitude with general deepening resulting from sea-level rise, transgressive facies changes, and increasing tidal amplitudes after wetland reclamation (Allen and Rae, 1988). In combination with sediment budget deficiencies, sea-level rise is a major risk (e.g. Regnauld and Fournier 1994; Shennan *et al.* 1999). On the short-term scale, sensitivity is shown to be particularly important with respect to variations in meteorological parameters such as precipitation (e.g. Provansal 1995a) and storms (Irion 1994; Regnauld *et al.* 1996; Wheeler *et al.* 1999), especially during winter (Duffy and Devoy 1999). Any changes in storm frequency/magnitude response resulting from climate change is likely to have a major impact on these environments, but again, can be compensated to a certain degree by a sufficient sediment supply, although this is countered by an accelerated sea-level rise.

Again, human influence is a major factor in environmental change (e.g. Paskoff 1994; Pirazzoli 1995). Dredging of navigation channels close to tidal flats and marshes, riverine embankments and dykes, and exploitation/reclamation of those areas leading to a decrease of the tidal basin area and hence an increase of the tidal prism, can be considered as dominant impacts on the sediment supply to the environments and leading to permanent flooding and degradation under accelerated sea-level rise (e.g. Moreira 1992). In estuaries and embayments where damming of up-stream rivers also controls sediment supply, degradation of flats and marshes is prone to occur in the near future under these conditions.

Estuaries and deltas
The evolution of estuaries and deltas has a direct relationship with climate forcing. Schumm (1969) has shown that in fluvial basins, there is a direct effect of temperature changes on the relation between mean annual runoff, mean annual sediment yield and sediment concentration, and mean annual precipitation. It appears that for a given annual precipitation, sediment concentrations increase with annual temperature, whereas for a given annual temperature sediment concentrations decrease with an increase in annual precipitation (Schumm 1969). The effects of this translate downstream into variable water and sediment input into estuaries (e.g. the Varde Å estuary, a unique example of a natural, undyked estuary on the Danish Wadden Sea coast, Bartholdy and Pejrup 1994) and deltas. In combination with accelerated sea-level rise, major hazards can be expected in these environments: drowning of estuary mouths and adjacent wetlands (e.g. Dabrio *et al.* 1999; Freitas *et al.* 1999; see also previous section); decay of delta bodies by coastal erosion.

Deltas are especially sensitive to variations in sediment fluxes, coastal erosion through sea-level rise, land subsidence, and variations in storm patterns and amplitude (e.g. Bondesan *et al.* 1995; Bruzzi and Provansal 1996; Suanez and Provansal 1996). The combination of these makes deltas one of the most vulnerable coastal environments of the world. As human occupation in these areas is as a matter of fact also very high, particular attention is needed from coastal managers to assess delta sensitivity to environmental changes.

As stated, estuaries and especially deltas, have been favourite sites for human occupation since prehistoric times (e.g. the Nile delta). Human impact therefore is huge and complex in both environments (e.g. Morhange *et al.* 1996). In estuaries, land use and dykes together with dredging and regulation of navigation channels, have profoundly changed the general morphology and geometry of drainage patterns and fine sediment sinks (e.g. Paskoff 1994; Van den Berg *et al.* 1996). Bypassing of fine sediments due to an increased ebb-dominated runoff, especially with the creation of single navigation channels (Paskoff 1994), thus forms potentially serious concerns in relation to the existence of wetlands along estuaries.

In deltas, erosion/sedimentation patterns are profoundly influenced by human occupation and land use, both in the delta area and in the hinterland (e.g. Paskoff 1994; Bondesan *et al.* 1995; Provansal 1995b), leading to extension of delta lobes by increased sediment input, as well as by decay through regulation and damming of fluvial channels, in addition to natural erosion by coastal drift (e.g. Maldonado 1972; Smith and Abdel-Kader 1988; Bruzzi and Provansal 1996). The sensitivity to an accelerated sea-level rise is particularly increased in deltaic environments by water or oil extraction, leading to an increased subsidence.

Cliffs

Sea cliffs are, by definition, erosional features at the land/sea boundary and as such deserve a special place in the assessment of sensitivity impacts (e.g. Lee 1995; Shuisky 1995). Like in all coastal systems, and perhaps even more so, cliff erosion becomes problematic when human occupation of its tops is threatened. On the other hand, sediments derived from cliff erosion are important sources for downdrift coastal sites. Typically, it is primarily the recreational use of cliff tops and the beach zones at their base, that have started the present concern about cliff erosion in the first place (Paskoff, 1994). Natural processes of erosion are furthermore increased more often than not by engineering protection measures along the shoreface, construction of resorts and roads, trampling of cliff tops, and heightening of the phreatic water table (e.g. by the irrigation of gardens, the use of water from septic tanks, or the draining of swimming pools; Kuhn and Shepard 1984). It is therefore in those recreational areas that erosion assessment studies have been done and protection measures undertaken with variable results (e.g. Dias and Neal, 1992; Psuty and Moreira 1992; Marques and Andrade 1993; Regnauld *et al.* 1993; Bray and Hooke 1995; Regnauld *et al.* 1999).

5.2 AN ASSESSMENT OF A POSSIBLE RESPONSE TO COASTAL FORCING

From the above, an assessment can be made of possible measures towards a protection of coastal environments against climate change impacts like an accelerated sea-level rise or storm frequency/magnitude changes. This assessment is given in table 5.2. Due to the high occupation level of most coastal areas of the world, it is often impossible not to use artificial measures in order to stabilise or protect settlements and economically vital centres of human presence (e.g. De Mulder 1994). However, strategies have been developed and have to be implemented towards a better and more sustainable coastal environment in most cases. In short, three main adaptive strategies can be defined towards this goal: the retreat of human occupation, the accommodation of changes, and protection measures against changes (Bijlsma *et al.* 1992, 1994). A fourth, highly debatable one, would be doing nothing. Also, retreat can be viewed as simply retreat or managed retreat. In particular, managed retreat options lead for the time being to unclear solutions about what should really be done and are expected to involve counter-productive political decisions.

Adaptive strategies have in particular been presented in 'The Common Methodology for Assessment of Coastal Areas to Sea-Level Rise' (IPCC 1991; Bijlsma *et al.* 1992) which was severely criticised by Kay *et al.* (1996) who concluded that "*The Common Methodology* has failed to meet two of its three main objectives, namely, that of becoming (1) a globally applicable method for assessing the potential future coastal impacts of greenhouse effects, and that of (2) enabling the development of a global picture of coastal zone vulnerability to greenhouse effects. The third objective, however, that of assisting coastal nations in planning to reduce the impact of future greenhouse effects has been partially met through the raising of awareness of the sea-level rise issue within coastal nations".

Direct response to climatic forcing is mostly conjectural and related in many cases to derived impacts on the geosystem (Table 5.2). Priority should therefore be given to these impacts on coastal sensitivity. Response strategies there should be a combination of the three main adaptive strategies defined above, after a thorough assessment of local characteristics of coastal processes and impacts.

5.3. THE MESHING OF THE PHYSICAL AND CULTURAL LAYERS IN THE COASTAL ZONE: THE ROLE OF COASTAL ZONE MANAGEMENT

5.3.1 Sea level rise and the cultural response: the basic problem for CZM

The threat of relative sea-level rise is not something new to the human condition. At all stages of human development there has been an ongoing battle between natural processes of coastal adjustment and human requirements of living in the coastal zone. Any response of natural processes can only be viewed as having associated impacts when seen through the filter of human consequences: without humans there are no problems associated with sea-level change. The principal human problem associated with sea-level rise is the degree to which a society can remain spatially fixed in a dynamic coastal zone that is spatially unstable. The traditional approach to the issue of dynamic change was a function of the available technology by which any society could protect its fixed position. Society has learnt that only through the development of a common will and purpose can such protection offer any chance of success: piecemeal protection has never been sufficient. Only when the spatial scale of protection is of a similar resolution to the spatial scale of the process has there been any likelihood of effective resistance. However, even effective resistance has at best only slowed the impact of the process. Rarely has protection worked without deflecting the impacts in space and time to elsewhere on the coastline. The human memory of this history of coastal failure is poor, and it has not prevented an emphasis on protection and resistance to inundation as a *leitmotiv* passed down the generations, exemplified by the ancient Dutch expression "*water wolf*" that describes the permanent threat from inundations. Protection is still likely to be the major issue with respect to the problem of defining cultural adaptation to what is envisaged as the result of accelerated rates of relative sea-level rise over future decades.

The major problem associated with how society can manage the coastal change of the future is the great uncertainty of such changes. In outline, coastal scientists can offer a macro perspective that accelerated atmospheric warming is likely to raise relative sea-level, and that increased relative sea-level tends to a disproportionate increase in shoreline retreat. Research on impacts and adaptation is often bedeviled by the questions: 'Impacts of what?' and 'Adaptation to what?' (Burton 1997). In addition, relative sea-level rise is not happening at the same rate worldwide and it is even possible to point to areas where sea level has remained stable despite global sea-level changes. The mechanisms behind these regional trends are poorly understood to say the least and may be associated with aspects of deep-ocean circulation and basin morphology. Global warming may also be associated with greater activity in ocean-atmospheric exchanges that help fuel increases in storminess. The conjunction of increased storminess and raised extreme water-elevations indicates the reduction in return periods of inundation events. In brief, the future will see the present coastline as a wetter and more energetic zone that will suffer morphodynamic and ecological change, and as a consequence force cultural change. The current concern is coastal science's inability to make predictions of what will happen in any detail without substantial attached uncertainty. Success in modelling future forcing scenarios relates to our ability to translate macro conditions to regional contexts and to handle the potential non-linearities in coastal responses.

Where does this leave the 'human consequences' if the physical layer is so poorly understood? Accelerated coastal forcing will lead to greater vulnerability in coastal communities. Economically developed coastal societies are based on a complex web of interconnecting ties of an economic, social, religious and political, in short, cultural nature. Such a multi-factorial basis to society may prove to increase vulnerability to coastal forcing. The interconnections of cultural life generate emotional-based values and belief filters that imbue the cultural landscape with substantial implicit meaning. These filters added together with explicit economic value of threatened infrastructure, increase the spatial inertia of societies and underlie the inability of many to move when threatened by physical change. Thus, increasing development means continuing fixed-societies that suffer increasing vulnerability to coastal forcing.

Historically, and the European Union's coastline is no exception in this respect, the general rule is that coastal societies suffer under coastal changes. However, it is important to differentiate between secular and instantaneous changes in relative sea level and how society tends to deal with them. On the one hand, instantaneous relative sea-level changes are viewed by economically-developed societies as essentially 'reversible' high-magnitude events (floods and even tsunamis), associated with a high-energy expenditure (waves and/or winds) that leads to a 'temporary' dislocation of services and resources. People are usually unaware of the potential upper magnitude of extremes in this category and the severity of associated dislocation to the point of total loss. Instantaneous relative sea-level changes are viewed as being absorbable by an economically developed society through the medium of market forces and fiscal instruments (insurance, government mediation and in the final analysis, personal loss). Extreme events are often accepted in economically developing countries as an inevitable consequence of living at the margin of a hazardous environment. This fatalism is not an active element of societies in EU countries and it is likely that any increase in event magnitude and response in the future will be matched by cries for government intervention and bailing-out, rather than mute fatalistic acceptance. This encapsulates a present problem of the coastal zone: the payment by all for the protection of the few at the coastal zone. It is unlikely that this can continue given the sheer scale of escalating costs associated with the growing demand for protection.

On the other hand, a long-term secular change in relative sea level is something that most societies view as a low-cost, marginal event that threatens only isolated elements of societies' activities. This is seen as unthreatening as it is imperceptible: how can a few millimetres per year of sea-level rise have any impact on the range and diversity of traditional life at the coast? Despite sea-level rise's insidious, pervasive and non-reversible nature for all cultural operations at the shoreline, society does not seem to view secular change as being more potentially damaging than instantaneous relative sea-level change. Unfortunately, society does not have a universal knowledge, understanding, or regard for the dysfunctional basis of this long-term trend that pressurises the spatial sequencing of cultural activity and forces human retreat from the shoreline. Estimates of coastal concentrations of population are varied and prone to myth status. However, Cohen *et al.* (1997) show that 37% of the world's population live within 100 km of the coastline. About 5% of that proportion would be directly exposed to serious risk from storm-surge flooding given a 1 m rise in relative sea level. Most of this population (c.100 million people) will be forced to migrate inland. This would be among the biggest mass-movements of human population yet observed in the world's history. Reducing or coping with this end result defines the very pressing need for sustainable coastal zone management on an international scale.

Coastal Zone Management (CZM) is concerned with facilitating adaptation in societies' activities that reduces their vulnerability to coastal forcing. Adaptation is difficult given societies' past and current reliance on technical fixes for problems. The very word "adaptation" suggests the need to work with the coast rather than to use interventionist protection approaches that, by experience, are short-lived in their utility. The fact that technical fixes do not appear to work easily or cheaply with coastal problems means that adaptation has to be related to causes rather than symptoms of coastal living. This means working with cultural perspectives where inertia is an entrenched characteristic. In the face of increasing susceptibility to forcing extremes, CZM will have to deal at the same time with both changing the basis by which society relates to the coastal zone, as well as with reinforcing the resilience of the cultural layer to match the physical basis. The difference between susceptibility and resilience, indexes the sensitivity of the coastline, a factor that will change over time. Within Europe, there is a range of (culturally defined) attitudes that prevent a universal adaptation to the consequences of relative sea-level change. This is the fundamental problem that has caused the ever-increasing delay in the ability of the European Union to offer any real programmed response to CZM issues. Coastal sensitivity is increasing year-by-year while governments prevaricate over effective and sustainable policies.

(*Acknowledgements* Michele Capobianco is thanked for his contribution to section 5.3.1)

5.3.2 What do we mean by Coastal Zone Management?

World wide, there is a range of human developments along the shoreline, claiming to be part of coastal management schemes. It is important to be able to differentiate between spurious and worthwhile schemes. To do so requires a generic standard for coastal zone management to be set. In any worthwhile CZM scheme there should be elements of the following five principles.

- *Sustainability* of human-induced coastal change. At a minimum, the physical-cultural stock and value of the coast should be stable on an inter-generation basis. Sustainability, as the key principle to good CZM, is only likely if the remaining principles are satisfied.
- *Shoreline planning* should be integrated into a coastal zone framework.

- *Competing demands* for coastal resources should be balanced across the wide physical-ecological-cultural set and context of conflicting demands.
- *Cohesion of decision taking* between all coastal parties at central, regional and local level.
- *Flexibility* of built and non-built responses to coastal 'demands'.

Experience shows that a wide spectrum of coastal zone activities claim to be part of a 'management' process so, inevitably, there are good and bad examples of management. It is rare, when coastal protection is the explicit purpose of the activity, that the management can be seen as sustainable and hence defined as good (Orford 1993). The best CZM is of an integrated nature (Integrated Coastal Zone Management (ICZM): Sorensen 1997) when the following elements are combined.

- A lack of overt physical or cultural sectorialism in the elements being managed.
- Coastal waters, coastline and coastal lands are equally included in the planning remit.
- There is a framework for decision-making concerning all scarce resources.
- Applied coastal research is undertaken to define good coastal practice.
- Education of all coastal users (stakeholders).
- Public outreach in decision-making.

The achievement of an integrated status is difficult because of its all-encompassing nature. There is a tendency to believe that ICZM somehow will replace CZM, but the basis of CZM is still needed within ICZM. The emphasis of integration in the latter is related to the communality of purpose and approach to be found between all stakeholders. Horizontal sectorial interests (e.g. intra-government differences) need to be broken down and emphasis needs to be put on the vertical integration of CZM purpose and method between grass roots and central government. This inevitably has to be from the grass roots up, but good ICZM identifies national /regional policies which encompass sustainable methods at the local level.

5.3.3 Why do we need CZM: the threats in Europe

The coastal zone is the spatial response focus to both physical and cultural pressures. The last decade has shown elaboration of the threat to the shoreline induced via accelerated global warming (Titus 1987; Houghton *et al.* 1996). Sea-level rise is now regarded as an actual threat rather than a potential one (Houghton *et al.* 1996; Leggett 1996) and its web of direct and indirect responses should be a major driver of CZM. Unfortunately, the emphasis of this threat has resulted in skewed CZM where protection *per se* has been the substance of much management response. The second set of pressures is that of the cultural encroachment from the landward side of the coastal zone and it is in this area where management appears to be least successful.

Culturally, there are two phases of economic development that have set the modern agenda for coastal use in Western Europe. First, the historic and contemporary economic development of the coastline as a resource and manufactured-commodity break-of-transport, i.e. the coastline is of no intrinsic consequence viewed only as an obstacle to economic development. However, and at the same time, the coastline has been an ever-expanding opportunity area for economic and political expansion, in particular in Europe since the Late Middle Ages (the colonial empires and the slavery trade being two of the developments that shaped most profoundly the face of the world today). Second, the historically determined wave of tourism and recreation that initially showed as a 19th-century phase of concentration of intra-national holiday resorts (superimposed on fragile and unstable coasts). This was followed by a mid-20th century international phase of tourist pursuit of sun, sea, surf, sand

and sex (S^5 factor), i.e. the shoreline is initially explicit in the development process but over time its centrality is lost under the momentum of tourist growth.

These two economic imperatives were and are potent causes of unconstrained physical growth in the coastal zone. For example, tourism is big business, generating 10% of Spain's total earnings, thus 54% of Spain's population now lives within 50km of the coast (Fischer et al., 1995). Without integrated CZM, such explosions of physical growth will compromise society's ability to retain a sustainable coastal usage.

5.3.4 ICZM uptake in Europe

It is important to recognise that these coastal pressures operate incrementally and are not always apparent. However, although it is clear that the coastal nations in Europe have been given the stimulus for coastal management, there are a number of stumbling blocks to activity that can be observed. Much of the recent coastal zone research has been directed to the most urgent and financially pressing problems (e.g. IPCC's concern for islands and island nations) and thus indirectly promoted an urgent protective reaction. Other international work is directed at the science of change rather than the management of change (IGBP's LOICZ: Pernetta and Milliman 1995), with much of the European Union's R&D being concerned with the same changes. It is noteworthy that the main statutory directives from the European Union concerning CZM have been related to ecological purposes where, for example, the Habitats Directive related to the enhancement of bio-diversity, SCA and SPAs have found a strategic role in coastal zone development (Huggett 1997). It is ironic that, though some of the strongest control on development of coastal areas comes from the ecological imperative, it is scarcely the essential tool that planners need for CZM. A successful ICZM programme cannot be built around a few protected habitats despite contrary claims (DOE (NI); 1995).

Sorensen (1997) indicates that there were 140 ICZM programmes in 56 out of 177 possible nations (1993), of which 56 programmes were in the USA, and 50 programmes were only feasibility studies. These numbers do not suggest that ICZM has made much impact in national planning since the heady days of the 1972 Coastal Zone Act in the USA. If we examine the take-up in the light of an innovatory wave model, it would appear that we are still trailing along the floor in a slow leading-edge innovative domain (Figure 5.3). The question of when (or if) accelerated take-up will operate is worrying. The present threat scenarios do not appear to be pressurising coastal management sufficiently, yet the present reaction gap of opportunity is rapidly diminishing as we approach the domain of definite accelerated global threats. We tend to think about the magnitude of the sea-level threat over the next 50 years, but that could be minor compared to what is possible over the century-plus scale (Wigley 1995).

The innovation wave of CZM is slow in moving through Europe. Generally, at the national level there has been slow reaction to the needs of CZM after the starting gun of the California Coastal Commission (1968) and the CZM Act of the USA (1972). However, since the 1990s, CZM has begun to make some impact on the declared policies of central government agencies, to the extent that there has been a rise in the number of European Union countries who now specify CZM intent in coastal planning (Figure 5.3). But one still needs to differentiate between an *avowed* intent of central governments and the *reality* of pro-CZM action at the grass roots. If we were to look at CZM implementation on a site-by-site basis, then undoubtedly the European Union would still be in the 'innovators' section of the curve (Figure 5.3) given the total length of the European Union's coastline of concern: CZM still has a long way to go inside Europe!

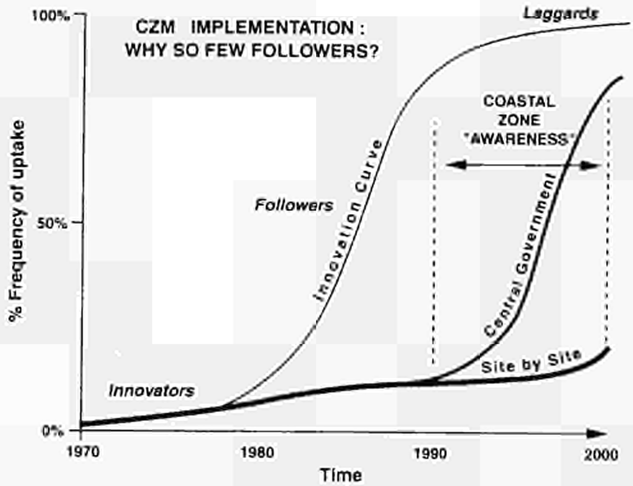

FIGURE 5.3 CZM take-up at both EU national and site-by-site level, as a schematic function of an 'innovation wave'.

If we look at the European Union's national response to coastal threats then, although there is contemporary action, there is also variation. One can observe some generic substance with protection as the initial and continuing stimulus to action. As the fiscal requirement to contain the coastal threat has become articulated (Kaufman and Pilkey 1982; Devoy 1992; UK National Audit Office 1994; Frankhauser 1995; Leggett 1996), the coastal nations have stumbled into a limited range of national attempts to rationalise their coastal situation, dictated to a large extent by the coastal typology. Although there has been much discussion concerning the four principal modes of coastal response: hold-the-line, accommodation, do-nothing and managed retreat, few nations show a creative use of any of these instruments, steadfastly relying on hold-the-line. Belgium (Charlier and De Meyer 1995), the Netherlands (Louisse and Van der Meulen 1991; Koster and Hillen 1995) and Denmark (Jensen 1994) identify national coastal programmes. The latter two focus on dune management as natural protection barriers, while in the Netherlands the policy includes continuous sediment nourishment on the beaches and the foreshore (TAW 1995a and b). Though claiming ecological soundness and reliance on 'soft' principles and hence sustainability, these programmes are essentially hold-the-line policies for protective purposes. The German North Sea coast with its long tradition of surge protection by major dykes or bunds, and the policy of engineering soft sediment deposition by bulkheads and palisades, is historically constrained into a protectionist mode. The United Kingdom has laid great emphasis on a regional coastal-cell-based approach to management of the coast through its Shoreline Management Plan (MAFF 1995), an approach that requires 'integrative' planning of coherent coastal units, but the protectionist mode is still high on the agenda in most plans. There is an identifiable mood that SMPs are a move towards ICZM . However, this perspective must be viewed as contentious given the lack of explicit cultural-layer emphasis in SMPs. Ireland is still committed to a protectionist stance, despite pleas from coastal county engineers for alternatives, given the counties perilous fiscal budgetary state for engineered solutions

(NCEC 1992). Spain in principle, has moved towards a proactive management stance by identifying a national version of set-back by declaring a 100 m coastal strip that should be free of development (Shores Act 1990: Fischer *et al.* 1995). However, easements are becoming common when claims for a (generally tourist-related) public infrastructure are made. This tourist pressure, causing the accelerated Mediterranean coastal development of the 1990s, has led almost axiomatically to a protectionist response. More disturbingly, Fischer *et al.* (1995) identify a lack of coastal knowledge in coastal planners, a problem that is likely to be wider than Spain alone.

5.3.5 Why so few ICZM programmes in Europe

Why are there so few ICZM programmes? It is probable that the available spurs to action are not the best ones or have not been articulated into adequate action statements. Existing European Union's CZM statutory pressures are insufficient: you cannot build ICZM around bio-diversity. Where national statutory CZM policy has evolved, then concerted action is feasible. However, in the case of the Netherlands where the most successful statutory policy is available (Water Defences Bill involving protection, fiscal and administrative instruments as well as educational ones) the outcome is not an ICZM *sensu stricto*, given their dominant protectionist mode. The greatest need for ICZM is where several sectorial pressures co-exist, yet experience shows that in this situation the pressure of the built-environment tends to dominate, leading to a protectionist stance which, over time, becomes self-fulfilling. As Doyle *et al.* (1987) stated, 'once you start engineering on the shoreline you are committed to a lifetime of engineering'. At this rate, a possible future scenario emerges where we will be dichotomising coastal management between two entities: protected-built (fiscally and populist driven) and protected-habitats (legally and minority driven).

There appear to be two distinct approaches to CZM: proactive and reactive modes of management (Table 5.3). Only a proactive approach will enable ICZM while the other (reactive) leads unfortunately to a *status quo* approach that is inevitably protectionist (Table 5.3). Some might argue that the two approaches are ends of a continuum in which middle ground is the best that can be achieved. But experience shows that there are few if any countries occupying the middle ground. It would appear that the coastal zone's greatest problem is how to move management towards a proactive stance, as national choice of CZM mode is severely constrained by a number of institutional filters which tend to work in support of reactive-management policies. Proactive management requires statutory process as few nations have existing legislation to contend with the requirements of ICZM. Governments do not universally welcome the need for a statutory authority or instrument, by which the whole coastal zone can be managed. It is rare for a concerted coastal zone body to be created with executive management powers. In many cases, central governments see this as anti-democratic and draconian (cf. the history of the Californian Coastal Commission). The United Kingdom has consistently turned its face away from this path, even going against the view of the House of Commons (1992) that a unified statutory agency was needed. It has argued a need for local inputs to shoreline management, despite the partisan attitude prevalent at local level as well as the distinct lack of scientific input to decision-making at this level. Likewise, the Irish government has resisted the call for a single statutory coastal agency. One can only wonder whether this resistance is really about avoiding anti-democratic decision-taking, or is more a case of existing government structures and bureaucracy with vested interests, wanting to avoid loss of power and resources. It is noticeable that where a central government agency's remit already covers the coast (e.g. the Netherlands), experience shows willingness for statutory changes.

Thus, the lack of statutory powers is the principal reason why proactive CZM does not occur. The viability of any policy other than *the status quo* is virtually non-existent. Coastal policy has usually to be worked out within the context of existing governmental structures and fiscal policies. Without cause for change, the coastal zone becomes the plaything of cultural systems that do not recognise the value of the coastal zone *per se*. Existing legislation is often fiscally active for built-protection but unavailable for purposeful intervention in a non-built-protection mode. Cost and inertia are joint vehicles for reactive management. The final filter, and despite being the most prominent, the least understood, is that of culture by which a society perceives, values, and uses the coast. Cultural tradition proves to be the strongest inertia to cultural coastal change

TABLE 5.3: Potential modes of coastal zone management

PROACTIVE / IDEAL • Management of *whole* coastal zone is central to policy. • *Sustainability* is central to policy. • *New legislation* to ensure statutory nature of coastal planning and directives. • Stress *continuity and coherence* of specific coastal systems. • Coastal protection is one of a *series* of policy approaches to coastal problems. • Sensitivity to *justification* of coastal decisions. • *Education* of coastal parties.
REACTIVE / PRAGMATIC • Use of *existing legislation.* • Existing legislation and governmental structures allow *coastal protection* to appear as the *central coastal management policy.* • *Minor* non-statutory planning changes. • Re-jigging of *existing bureaucratic/governmental structures.* • Decision-making left at *local level* supported by voluntary ad-hoc groupings of coastal communities (e.g. the 'rich cascade' of English coastal cells). • *Splitting of control* on elements of otherwise coherent coastal systems by governmental administrative boundaries. • *Partisan* grass roots decision-making as an attempt to meet democratic deficits of central government.

5.3.6 Steps to implementing (I)CZM in Europe

The variability of Europe's coastline indicates that CZM experience based solely on an east-coast USA model of barrier islands is not sufficient. We need to develop coastal models appropriate to European coastlines. The physical dynamics of the European coastline are still relatively unknown especially when considered on a time scale of decades or centuries (mesoscale). We have more to understand about mesoscale sea-level variability and storminess changes, which will cause non-linearity of process and shoreline response with associated thresholds of change (Leggett 1996). The complex interactions of coastal forcing and coastal response require us to know more about coastal sensitivity through the associated issues of coastal susceptibility and coastal resilience (Bijlsma 1996). Sensitivity needs to be set initially in a context of coastal type, yet we are still short of a full dynamic typology of Europe's coastline. We know little of the cultural framework of coastal living and usage around the European Union's coastline. Nor do we, given the environmental threats, understand how risks in a hazardous environment can be accommodated within the market-oriented economies of the European Union. Would communities adapt to temporary occupancy or do we have an economic free-for-all? It is clear that CZM cannot be divorced from the national government infrastructure in which CZM is implemented and controlled. The last decades of subsidiarity, free-market experience and devolution of government regulation, suggests that unitary coastal authorities with statutory power are even less likely in the next decade. This means that within the European Union, unless central ICZM directives are forthcoming, ICZM has to move towards a local process of educating the principal stakeholders in coastal affairs. Inevitably this will be a piecemeal process in which specific context rather than generic science takes a high profile. This will not augur well for the long-needed generic basis to ICZM development and implementation within the European Union.

5.4 FUTURE EU RESEARCH REQUIREMENTS FOR THE SUCCESSFUL IMPLEMENTATION OF (I)CZM

It is hardly possible, nor is it recommended, to draw-up general rules for coastal sensitivity to climate change on the basis of present knowledge of coastal processes. Especially temporal and spatial extrapolations should be used with great care (Figure 5.4; de Groot 1999). To this end, future sensitivity/vulnerability assessment methodologies should not be too prescriptive and inflexible like the IPCC Common Methodology (IPCC 1991). Alternatively, broad, flexible, and non-prescriptive guidelines may result in low levels of direct applicability. Therefore, a vulnerability assessment framework is recommended as the best way to integrate vulnerability assessment within national planning processes and strategies for anticipating and mitigating the impacts of climate change in the coastal zone (Kay *et al.* 1996; IPCC 1993), but in studying the adaptation to extreme events and the adaptation to climate change, the research communities should work more closely together (Burton 1997).

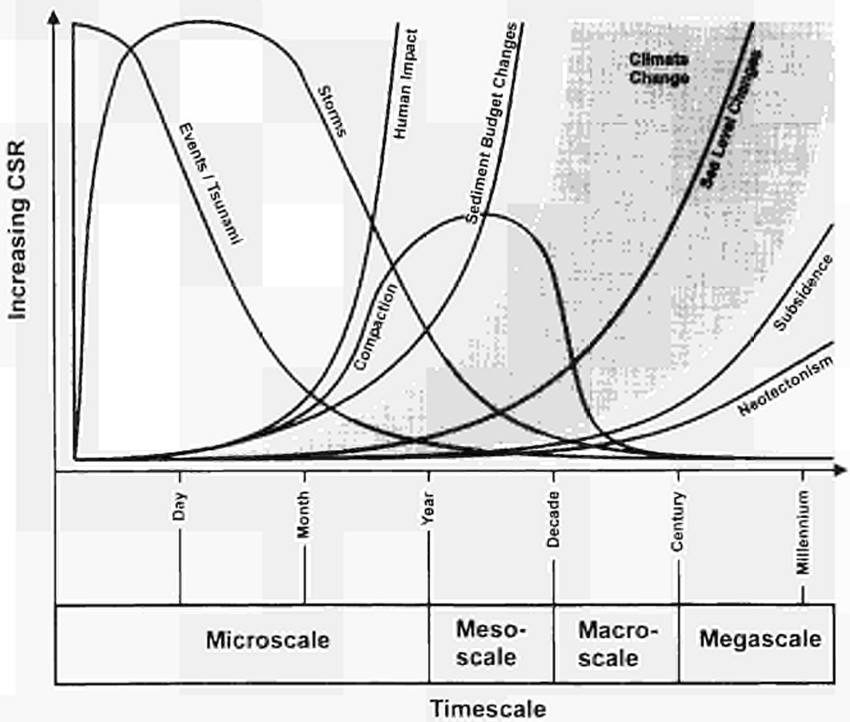

FIGURE 5.4 Coastal Sensitivity Response (CSR) of an array of forcing factors through time (de Groot 1999).

The metric of coastal sensitivity remains a challenge for the future. The capacity of communities to live with the physical sensitivity of their coastline indicates their vulnerability. But for a central government to respond to this as a national issue requires specifications based on common metrics. The contemporary debate on sensitivity tends to be semantically driven. Coastal science needs urgently to move beyond definitions and into consistent measurement. Sensitivity must reflect the cultural realm as much as the physical one. Hence, research must be undertaken into understanding the basis of cultural use of the coastal zone and how cultural interdependencies can be made more resilient by adaptation in the face of environmental instability.

Response strategies should be mainly concerned with the direct impacts of coastal changes, whether or not they are forced by climatic shifts. A combination of retreat, accommodation and protection adaptive strategies are commonly considered to be the best answers to environmental changes. The human impact should be answered in the same way, even if this means a set-back in current environmental or economic development of particular zones at risk. Mitigation techniques that closely replicate natural processes are to be preferred in all cases. Potential tsunami impacts are one of the few hazards that are difficult or impossible to protect from, but should be taken into account in coastal zone management assessments.

The lack of a coherent generic science to coastal decision-making underlies the problems faced in research programmes concerned with CZM. Although coastal science has developed a predictive capacity at the instantaneous scale, it is still far short of doing the same at the decadal scale. It is essential that the nature and structure of coastal change at different time scales be understood. The basis for these approaches is likely to shift from the physical dynamics of the instantaneous model to the behavioural orientation of the longer term model (Capobianco *et al.*, 1999). Models are likely to be regionally dependent and spatially defined in terms of wave-sediment interaction cells. Coastlines in the European Union cover a wide range of geomorphological and biogenic types that need to be explored for their patterns of development and of change. Such change is not always linear and the generation of non-linearities between coastal interactions of energy, sediment and spatial context ensure, particularly in the light of long-term sediment supply variation, the possibility of sudden shifts in the morphodynamic domain. The boundary conditions to such shifts need to be understood, especially given the impact of protectionist measures on sediment supply continuity to the shoreline. In particular, the implications for the spatial shift of shoreline components during domain change, remains a challenging issue that is at the heart of the problem for coastal communities. The use of set-back as a dynamic tool for management is impractical without an understanding of the spatial rates associated with coastal change.

What then, should be the generic basis for an *integrated* coastal zone management as a response to accelerated sea-level rise? Calls for action have been formulated recently, deriving from the EUROCOAST-Littoral '96 conference (Taussik and Mitchell 1996; Taussik 1997). First of all, the clear assessment of the zones at risk and the economic, biological, and geomorphological implications. In many cases, this has been done although follow-up is lagging behind. Secondly, and to support assessments, permanent multidisciplinary coastal observatories should be established. Thirdly, proactive scenarios rather than simply defensive ones, should be developed for the threatened coastal zones. These scenarios most probably will encompass the definition of set-back zones and mitigation techniques best appropriate for each particular area. Policy statements regarding these scenarios should be SMART (Specific, Measurable, Agreed, Realistic, Time based; Taussik 1997) and the long-term responsibilities and commitments of all partners to the plans should be clearly stated. There is a clear need in addition for awareness raising, demonstration of benefits, and the establishment of targets and indicators of sustainability through the dissemination of research and the commitment to established indicators. Fourthly, implementation of the best scenario for each particular coastal area should be driven through.

From the onset, a number of major problems arise from this generic view. First of all, many coastal zones are not supervised by single national entities. Instead, a large array of national and local authorities claim the responsibility to small stretches of the coastline. Co-operation and financial flow between these different bodies is more often than not hampered by legislation and economic/cultural barriers or interests. Lip service is paid to European Union or even national regulations, but measures are often difficult to implement because of these barriers and interests. Secondly, the economic interests and uses of the coastal zone are predominant on the local level. Few authorities, possibly none at all, are willing to become politically unpopular by enforcing draconian set-back measures on their coastal territories, especially if those measures imply a substantial economic decline. In the end however, it is the national governments that are responsible for the unavoidable economic and natural losses likely to result from an accelerated sea-level rise if nothing is done.

A number of salient points have emerged from both past and ongoing European Union research programmes, as to where future research in support of the issues raised for CZM by rising sea-levels should be directed. The need for a coherent research design that specifically addresses the issue of CZM in a science-based approach, rather than the contemporary *ad hoc* applied-approach dependent solely on engineering practice is awaited. The reason for the latter state of affairs is clear. The issue of protection against rising water levels is paramount in coastal communities and the responsibility for dealing with this lies in the hands of engineers concerned with site-specific problems, but limited in the remit of approach. The role of coastal science at the coastline is still full of uncertainty and the gap between what is view in theory and what can be put into practice is still too large to ensure implementation of long-term solutions as identified through science.

One of the blocks to such solutions lies in the cultural realm. The lack of understanding at grass roots and the lack of transmission of ideas and knowledge between science and community are a major concern. Transmission can only take place between science and community when value systems are consistent. The dissonance between science and community values over environment and its worth is variable around the European Union. Why value-differences occur and how the meeting of minds can take place are universal problems, their involvement in the coastal zone is no exception. However, there is an immediacy of rising seas that means this issue has to be approached and pathways to mediation have to be developed quickly. The current EU-LIFE demonstration programmes concerned with CZM offer one such solution to this problem (see also: European Commission 1999). There is a danger that such programmes will degenerate into site-specific resolution of immediate threats by protection rather than considering alternative perspectives that are dependent on the coastal education of the local decision-makers. The support of coastal education at local and national level should be an associated aim of any European Union programme related to CZM decision-making as well as decision taking.

Challenges and opportunities for the earth science community are manifold. In order to provide useful answers towards an integrated coastal zone management, a number of research questions have to be answered through them (Williams *et al.* 1991). These questions address problems of coastal processes (e.g. wind, wave, tide, and seasonal weather pattern impacts;

sediment sources, volumes and movement rates; short and long term processes; biologic communities), short-term coastal events (e.g. storm frequencies; storm surge impacts; erosion/sedimentation changes; permanent vs. temporary impacts), long term changes in water levels (e.g. paleoclimatic variations; paleo-sea-level changes; local versus regional scale impacts; future effects), and human involvement along coasts (e.g. groundwater, construction materials, and mineral resources; engineering structures; mitigation techniques; impacts on infrastructural works; river damming; fluid and mineral extractions; contaminants). Only through the answer to these fundamental questions can an integrated coastal zone management leading to a sustainable environment be effected.

There should also be a continuing exploration of the diversity in coastal zone methodologies. The use of non-built approaches to the management of coasts needs to be developed as soon as possible. However, it is clear that without 'coastal education' these approaches will not meet with local approval, regardless of their necessity in terms of environmental and, ultimately, societal need. This raises the fundamental requirement underlying any successful implementation of ICZM, that of educating all the stakeholders in the realities of living with the coast in the next millennium. This will be a major task since without consideration of its delivery, ICZM of the future will be no more than protectionist-driven and an unsustainable expensive sink of scarce-resources.

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Luxembourg: Office for Official Publications of the European Communities

2000 — XI, 247 pp. — 17.6 x 25 cm

ISBN 92-828-9023-6

Price (excluding VAT) in Luxembourg: EUR 38.50

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