

Electronic Supplement S3 to Chapter 3 Past and ongoing changes in the North Sea and its interface regions

This Electronic (E-)Supplement provides details on the evidence for variability in temperature (S3.1), salinity (S3.2), circulation (S3.3), coastal erosion and offshore morphology (S3.4).

S3.1 Temperature: evidence

Temperature is the mostly widely measured variable; evidence is relatively abundant, especially by using models constrained by observations. Sea-surface temperature (SST) can be measured *in-situ*, and by satellites from which data typically need adjustment for changing atmospheric composition, using *in-situ* data (e.g. Advanced Very High Resolution Radiometer – AVHRR – are adjusted, Along-Track Scanning Radiometer – ATSR – are not). Accuracy thus depends on *in situ* data quality and representativeness of local conditions. Infra-red sensors cannot measure through cloud; microwave sensors can (e.g. Advanced Microwave Scanning Radiometer; AMSR) but with less resolution and quality, and not near coasts. Errors are equivalent to several years of climate trend.

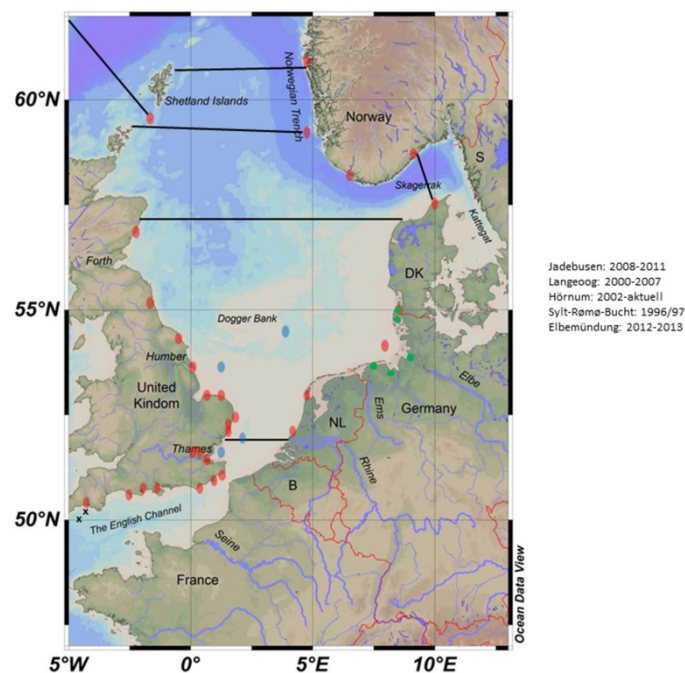


Fig. S3.1 Map of coastal monitoring stations (*red ovals & green circles*), offshore Smart buoys (*blue circles*), western Channel stations L4 and E1 (*crosses*), repeated hydrographic sections (*lines*)

Many (near-) coastal stations measure temperature, usually salinity and often other meteorological and oceanographic variables. Many (not all) locations are shown in Fig. S3.1. Sources of information include FRS (2007; now Marine Scotland Science) for Scotland, Cefas for England, van Aken (2010), van Beusekom et al. (2009), Flöser et al. (2009), Wiltshire and Manly (2004), Hickel (1998), Albrechtsen et al. (2012). Marsdiep SST dates from 1861; most series span some decades to the present. Offshore (Fig. S3.1) there are four Cefas “Smart” buoys in the North Sea or Thames estuary (from 2000 or later). Western Channel stations E1 (in 75m) and L4 (near-coastal) span over a century (Fig. S3.2; Smyth et al. 2010).

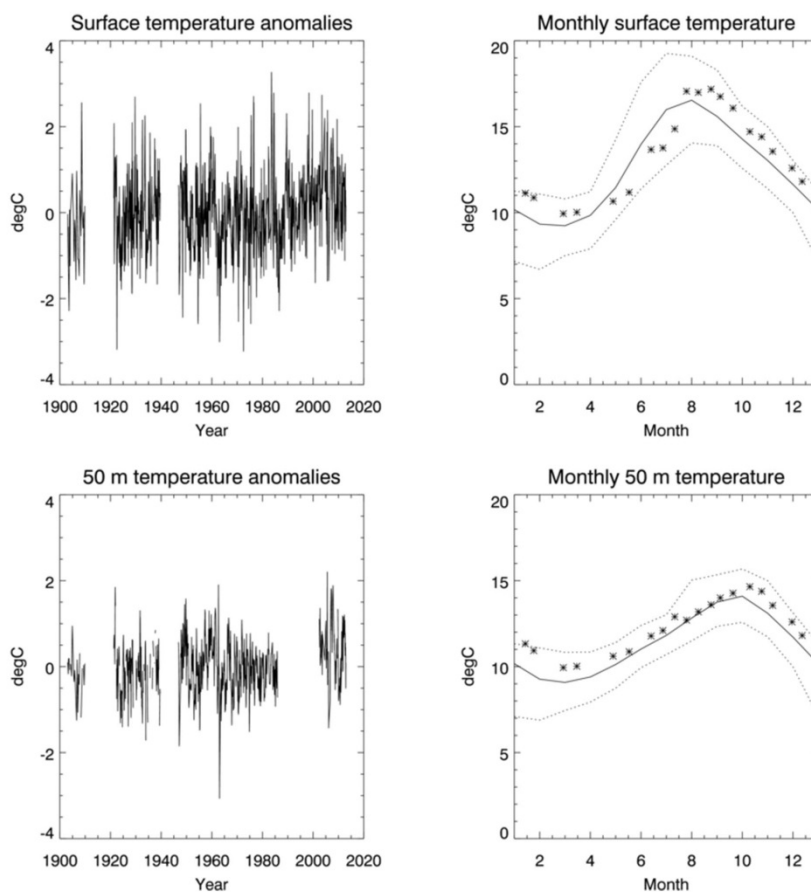


Fig. S3.2 E1 temperature anomalies (*left*), monthly averages (*right*). *Top panels*: surface layer (about 2 m). 1987-2002 data (*upper left*) are monthly averaged satellite SSTs from the Pathfinder dataset based on AVHRR (<http://www.nodc.noaa.gov/sog/pathfinder4km/userguide.html>). *Bottom panels*: 50 m depth. Asterisks (*right panels*): 2012 data; *dotted lines*: mean \pm 2 standard deviations. (Smyth et al. 2010, updated)

Ferries and cargo ships have allowed frequent monitoring of temperature and salinity since 1970, with “FerryBoxes” since 2002 (Petersen et al. 2009). Ship and buoy SST may provide *in situ* data to calibrate/validate space-based measurements which give more coverage. In the German Wadden Sea a number of stations have operated since 1996 during summer months measuring, among other variables, temperature and salinity (Flöser et al. 2009).

However, observations of temperature well below the surface are relatively sparse in time (surveys or sections, often years apart) or in space (moorings, typically 100s km apart). A long-term section is Fair Isle – Munken (across the Faroe-Shetland Channel) since 1900, now 3-4 times per year. This and others spanning several decades are shown in Fig. S3.1. Sources of information are Sherwin et al. (2012), Kangas et al. (2006), Albrechtsen et al. (2012). Profiling “Argo” floats data greatly improve estimates of ocean temperature and salinity variability down to 2 km and so help North Sea model estimates. The International Council for the Exploration of the Sea (ICES) co-ordinates winter surveys (IBTS – International Bottom Trawl Survey) providing annual winter bottom temperatures (and salinities) in the whole North Sea. In summary, sub-surface spatial resolution is sparse relative to scales of some phenomena (e.g. fronts and plumes) and annual or shorter periods are rarely resolved.

There are a number of attempts to derived gridded data sets (Table S3.1). They rely heavily on interpolation and different reconstruction methods and become increasingly uncertain for earlier years. For example, bias corrections compensate for changes in SST measuring practices before 1942 (c.f. main text Sect. 3.2.1). Calculated uncertainties (associated with measuring practice) accompany these data (e.g. Rayner et al. 2006; except HadISST). HadISST and ERSST interpolate using empirical orthogonal function (EOF) mode reconstruction and may not represent local conditions well. NOCv2.0 should represent SST well in well-sampled regions (*via* optimal interpolation with spatial smoothing scale 300km). In the North Sea, Advanced

Microwave Scanning Radiometer (AMSR) data reduce error estimates and alter details of the NOAA 1/4 degree dataset. Uncertainties in average gridded SST estimates are of order 0.2°C; typically larger near coasts and less in the central North Sea. However, such gridded data are specifically for the surface and show less skill for finer scales in time (less than a few years) and space (where local conditions may be influential; Hughes et al. 2009).

Table S3.1 Gridded data sets

Dataset	Period	Resolution	Reconstruction?
HadSST3 (Kennedy et al. 2011a,b)	1854-	5° monthly	No
HadISST (Rayner et al. 2003)	1854-	1° monthly	Yes
ERSST v3b (Smith et al. 2008)	1854-	2° monthly	Yes
NOCv2 (Berry and Kent 2009, 2011)	1971-	1° monthly	No
NOAA OI 1/4° (Reynolds et al. 2007)	1981-	0.25° daily	No, but uses modes to adjust bias
NOAA OI 1/4° with AMSR	June 2002 – Oct. 2011	0.25° daily	No (ditto)
KLIWAS North Sea Climatology	1890-2011	179 depth levels, 0.25° x 0.5°, yearly & monthly	No

Offshore observations in the North Atlantic and Nordic Seas are summarised annually in the ICES Report on Ocean Climate (Holliday et al. 2009; Hughes et al. 2010, 2011; Beszczynska-Möller and Dye 2013). Specifically for the North Sea, the Bundesamt für Seeschifffahrt und Hydrographie (BSH) has produced weekly SST maps since 1968 and monthly heat-content maps (from numerical models constrained by data): http://www.bsh.de/en/Marine_data/Observations/Water_temperatures_and_heat_contents/index.jsp.

1902-1954 average distributions of SST and near-bed temperature are discussed in Lee (1980), and their 1971-2000 climatology in Berx and Hughes (2009). Long-term trends of annual-mean North Sea temperature are shown by, e.g., EEA (2012; based on BSH and MyOcean data), MCCIP (2010) and HadISST1.1 (UK coastal waters, www.metoffice.gov.uk/hadobs).

Products for temperature from models can combine observations with dynamical constraints for a best estimate. For example, SODA (Haekkinen et al. 2011) assimilates available hydrography with forcing by European Centre for Medium-Range Weather Forecasts (ECMWF) winds from 1958. She et al. (2007) assimilated satellite and *in situ* data in a North-Sea-plus-Baltic model and used independent data to find a yearly mean model bias 0.07 °C and rms error 0.64 °C. Satellite data (calibrated by *in situ* SST) dominate the observational input. ECOHAM4 simulated SST for 1970-2006, forced by the US joint National Centers for Environmental Prediction (NCEP) and National Center for Atmospheric Research (NCAR) reanalysis (Kalnay et al. 1996) without data assimilation, showed a model bias O(0.5 °C) relative to observations (BSH compilation); the bias was attributed to heating only the upper model layer (Lorkowski et al. 2012).

Meyer et al. (2011) performed a 3D baroclinic hindcast of the North Sea (to 61°N, 5°W to 14°E but only to 2°W in the Channel) for 1948 to 2007, using HAMSOM (resolution 20 km × 20 km × 19 levels) forced by the NCEP/NCAR reanalysis. Holt et al. (2012) studied temperature of the northwest European shelf seas for 1960–2004 using (i) POLCOMS with ERA-40 surface fluxes, (ii) *in situ* multi-annual time-series, (iii) satellite SST, (iv) the ICES data base. All (i) – (iv) give a consistent picture for trends and variability. POLCOMS temperature root-mean-square error over one annual cycle (2001-2) was just over 1°C (Artoli et al. 2012). The EU “Global Monitoring for Environment and Security” (GMES) project MyOcean has weekly sea temperature reanalyses, e.g. for the north-west European shelf with SST data assimilation (O’Dea et al. 2012).

S3.2 Salinity: evidence

Salinity evidence is sparse relative to that for temperature; often too sparse to define smaller-scale influences (e.g. river plumes) or any seasonal cycle accurately. Measurements are at most of the (near-) coastal stations in Sect. S3.1, along with other observations described there: Western Channel stations L4, E1 (Fig. S3.3), ferries, cargo ships (and “FerryBoxes”), long-term sections, IBTS covering the North Sea annually, Argo profiling floats. Calibration is good, especially since 1970.

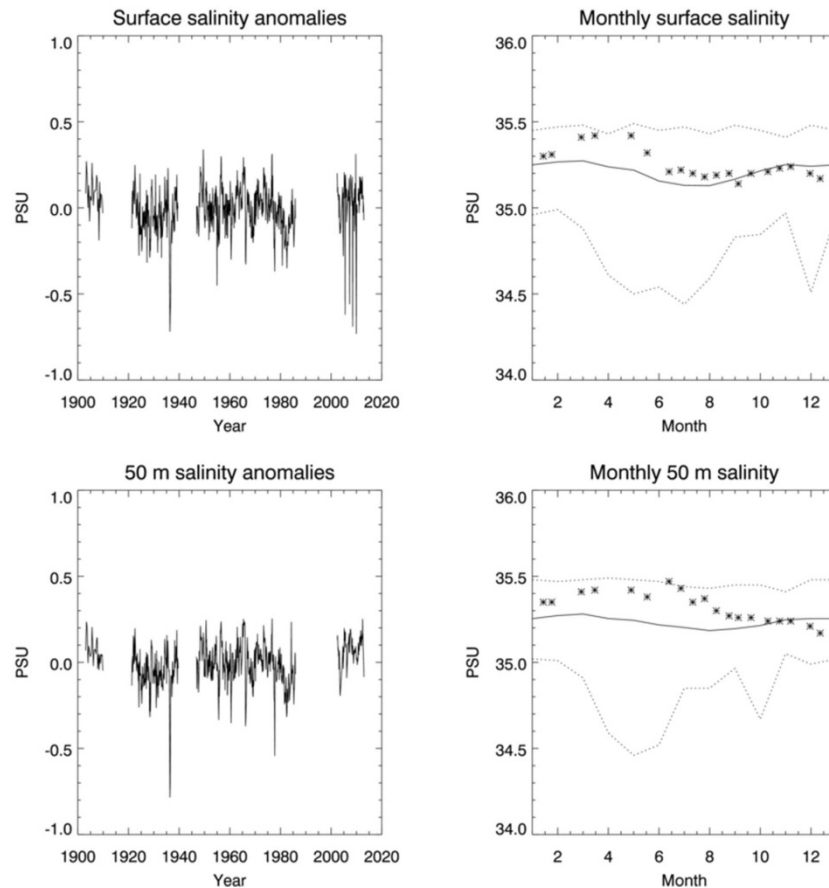


Fig. S3.3 E1 salinity anomalies (*left*), monthly averages (*right*). *Top*: surface layer (about 2 m). *Bottom*: 50 m depth. Asterisks (*right*): 2012 data; *dotted lines*: mean \pm 2 standard deviations. (Smyth et al. 2010, updated)

Damm (1997) compiled North Sea seasonal salt content and Berx and Hughes (2009) present a 1971-2000 climatology of surface and near-bed salinity. North Atlantic and Nordic Seas observations are summarised in the annual ICES Report on Ocean Climate (Holliday et al. 2009, Hughes et al. 2010, 2011; Beszczynska-Möller and Dye 2013). Modelled salinity over one annual cycle (2001-2; POLCOMS, ~ 12 km resolution) had RMS error 0.75, much larger than Atlantic changes (Artoli et al. 2012) especially near river discharges.

S3.3 Circulation: evidence

The Atlantic meridional overturning circulation (AMOC) can be estimated from hydrography in ocean-basin-wide sections; a 26°N section was surveyed five times from 1957 to 2004. Moorings to give time series across 26.5°N began in 2004 in the project “RAPID” (now “RAPID-Watch”; <http://noc.ac.uk/science-technology/climate-sea-level/rapid-watch>; Rayner et al. 2011). AMOC has also been associated with thermohaline forcing (Grist et al. 2009) to give reconstructions based on surface fluxes of heat and freshwater (Josey et al. 2009). Model hindcasts and data assimilation are discussed in Cunningham et al. (2010) and show promise for estimating AMOC natural variability.

Long-term circulation in shelf seas has been inferred from distributions of ‘tracers’ such as salinity and radionuclides, for example, caesium-137 and technetium-99 (Prandle 1984; Dahlgard 1995; Kershaw et al. 2004). Tracks of drifters and floats have also been used (Hill et al. 2008). Moorings on the West Shetland slope monitor Atlantic water transport past northern Scotland (Berx et al. 2013). Satellite altimetry may give

variations in surface currents. HF radar has been used temporarily in several areas around UK coasts [and for some years in Liverpool Bay] as well as the German Bight (Port et al. 2011). Except for HF radar which is limited in extent, all methods of measurement are sparse in practice. The numbers of current meter moorings represent much effort, but by decade the area-averaged spatial resolution is at best about 20 km; this is much coarser than the spatial and temporal scales (days) of shelf-sea eddies.

This emphasises use of numerical hydrodynamic models (Huthnance 2010). Tide-surge models are good for estimating wind-driven currents on time-scales of several hours to one day; this is important given the large contribution of wind-driven “events” to net circulation (main text Sect. 3.3.2). Holt et al. (2005) assessed uncertainties in a 3-dimensional baroclinic circulation model (POLCOMS) of the northwest European continental shelf. With ~ 7 km horizontal resolution and 20 ‘s-levels’ in the vertical, tidal currents were well modelled, with root mean square (RMS) errors of less than 0.4 standard deviations. However, residual current speed had RMS errors similar to the standard deviation of the data for the period simulated (August 1988 to October 1989). Net west-to-east flow in the bulk of the Channel has been modelled (Salomon and Breton 1993; Salomon et al. 1993) and confirmed from distributions of radionuclides (Guegueniat et al. 1995).

Transports from models of the North-West European Shelf are contributed routinely by centres in the North-West European Shelf Operational Oceanography System (NOOS), a regional alliance associated with EuroGOOS. NOOS daily publishes net water transports across key sections in the region (e.g. Fig. S3.4); some centres’ data are available for nearly a decade, e.g. BSH-modelled North Sea - Baltic exchanges through the Danish Straits from 2004 (www.noos.cc). The EU GMES project MyOcean gives 3-D analysis products.

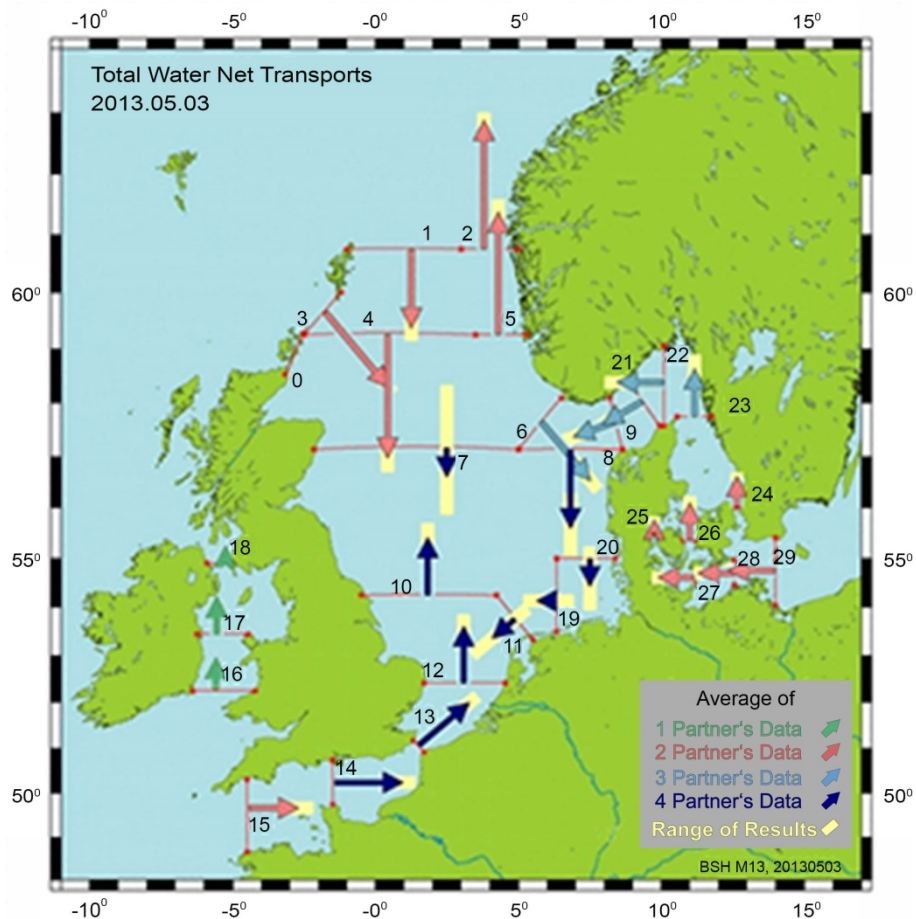


Fig. S3.4 Transports from an ensemble of up to four operational models from the NOOS (www.noos.cc) transports project

Fig. S3.4 also indicates general agreement between the flows predicted by the different operational forecast systems, despite different underpinning ocean modelling systems and different atmospheric forcing; this gives some confidence in the quality of the transports provided by NOOS.

Charting Progress (Defra 2005) tabulates mean transport across sections in the North Sea for the period 1987-1993 from three numerical models. Long-term 3-D simulations have been run: NORWECOM/POM for 1985-2007 (10 km resolution; Hjøllø et al. 2009) and for 1955-2008 (20 km resolution; Albretsen et al. 2012); POLCOMS for 1960-2004 (12 km resolution; Holt et al. 2009); HAMSOM for 1948 to 2007 (Meyer et al. 2011).

S3.4 Coastal erosion and offshore morphology: evidence

Coastal erosion and passive drowning, the two elements of landward coastline migration and land loss, take place on various time scales. Their long-term effects, most relevant in the context of climate change, are best captured by measurement time series spanning several decades or longer. Observation methodologies differ for different types of coasts and for different coastal environments. They focus on features that define the coastline at the time of measurement. As all of these indicators are dynamic, it is difficult to compare, harmonize and merge measurements, even if they are made by the same instrument and under similar conditions. This is especially true for the highly dynamic intertidal zone; the waterline moves back and forth in every tide.

The most suitable coastline indicators are easily and consistently identifiable, in the field and by remote sensing (Boak and Turner 2005). Morphological features include the crests and toes of foredunes, bluffs and cliffs. Not all of these features have the same indicative value. A cliff base, for example, can migrate seaward by 1m to 2m, whereas a cliff top only moves landward (see Brown 2008). Non-morphological features include high- and low-water lines. The above morphological indicators are located in the 'dry' intertidal and supratidal sections of the 'coastal tract', the cross-shore sequence of morpho-sedimentary sub-units as defined by Cowell et al. (2003). Areas below the low-tide line are also relevant to coastal-erosion studies and inventories, actively exchanging sediment with beaches, dunes and other 'dry' environments. This 'wet' section is generally monitored using echo sounders, but recently it has become possible to distil shallow subtidal bathymetry from video data at high temporal and spatial resolution, using time-exposure images.

Indicators above the low-tide line are monitored by a more diverse set of field and remote-sensing methods, each with its own capabilities and limitations (Moore 2000). Remote-sensing techniques are cost effective, reduce manual error and remove the subjective approach of conventional field techniques (Maiti and Bhattacharya 2009)

In profiling, the relative locations of points along the coast are determined by measuring horizontal distances, differences in elevation and directions, generally perpendicular to the coastline. Automated total stations, satellite-based RTK-GPS, and transit and plane table surveying can all be used. Shorelines are generally derived by interpolating between discrete profiles. (Large) distance between profiles limits the accuracy of interpolation (e.g. Cambers 1976; Pethick 1996).

Many coastal areas have extensive aerial-photo or satellite-image coverage, providing a valuable record of shoreline position. Photogrammetry creates digital terrain models from overlapping aerial photographs or satellite images. The interpretation of shoreline positions from aerial photography is subjective given the dynamic nature of the coastal environment. The highly variable high-water line is the most commonly used shoreline indicator because its associated colour change is visible on colour and on grey-scale aerial photographs and can be ground-truthed in the field (Leatherman 2003; Crowell et al. 1991). Terrestrial laser scanning and airborne LIDAR are used to build digital terrain models from which indicator elevations of morphological breaks can be determined.

In the event that a study requires the shoreline position to be constrained for times before the development of aerial photography, or if the location has poor photograph coverage, it is necessary to employ historical maps (Moore 2000). Scale differences, datum changes, distortions from uneven shrinkage, stretching, creases, tears and folds, different surveying standards, different publication standards, and projection differences are all sources of error (Boak and Turner 2005). Coastal maps and charts from before 1750 are generally much more inaccurate than later ones. For most coastal sites, the availability of data is so limited, however, that

low-quality datasets cannot be discarded (Boak and Turner 2005). To constrain former coastline positions lacking mapping and monitoring data, geomorphological proxy evidence may be used. Truncated landforms like foredune or beach ridges may be used to deduce locations and rates of past coastal erosion (e.g. Fig. S3.5).

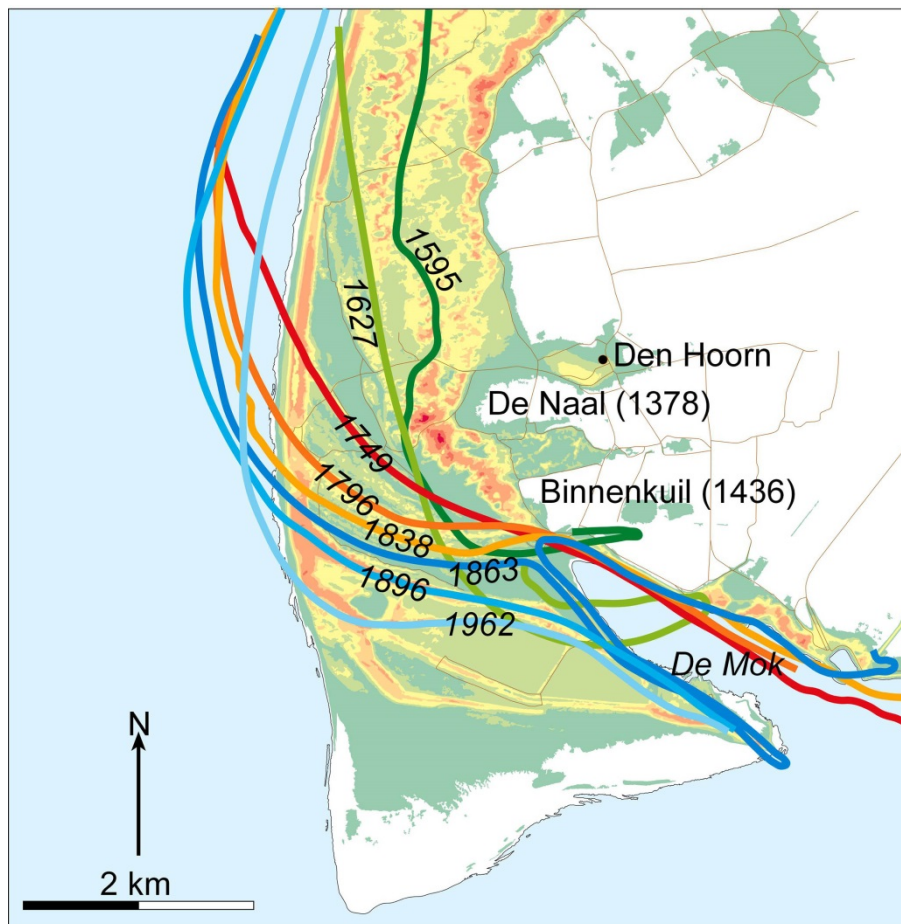


Fig. S3.5 Historical changes of southwest Texel. Coastal erosion, as inferred from the truncated dune ridges, is confirmed by the historical maps used to determine former coastlines (van Heteren et al. 2006)

Across Europe, large differences exist not only in the state of knowledge but also in the compatibility of data from multiple sources and in the accessibility of data and data products. Finding, comparing and merging data from echo-sounding, topographic surveying, photogrammetry, laser altimetry, historical maps, aerial photographs, satellite images and video cameras is not straightforward. For example, most main coastline indicators show temporal variability even within a single tidal cycle.

No single procedure translates one type of observation to another. Inconsistencies may be associated with (i) time of measurement: HWL varies between tidal cycles and also depends on the weather; (ii) methodology of observation: indicators measured directly and instantaneously will differ from those observed remotely or derived indirectly; (iii) indicators lacking accurate definition: for example, a dune foot may be taken as the break in slope at its base or as the beach-side edge of its vegetation.

Harmonization of coastal-monitoring data (e.g. Fig. S3.6) is generally based on assumptions regarding value of an indicator proxy, rather than on systematic use of correlations between different methodologies' time series. This assumption-based approach is good enough for broad, regional studies but inadequate for process-oriented case studies.

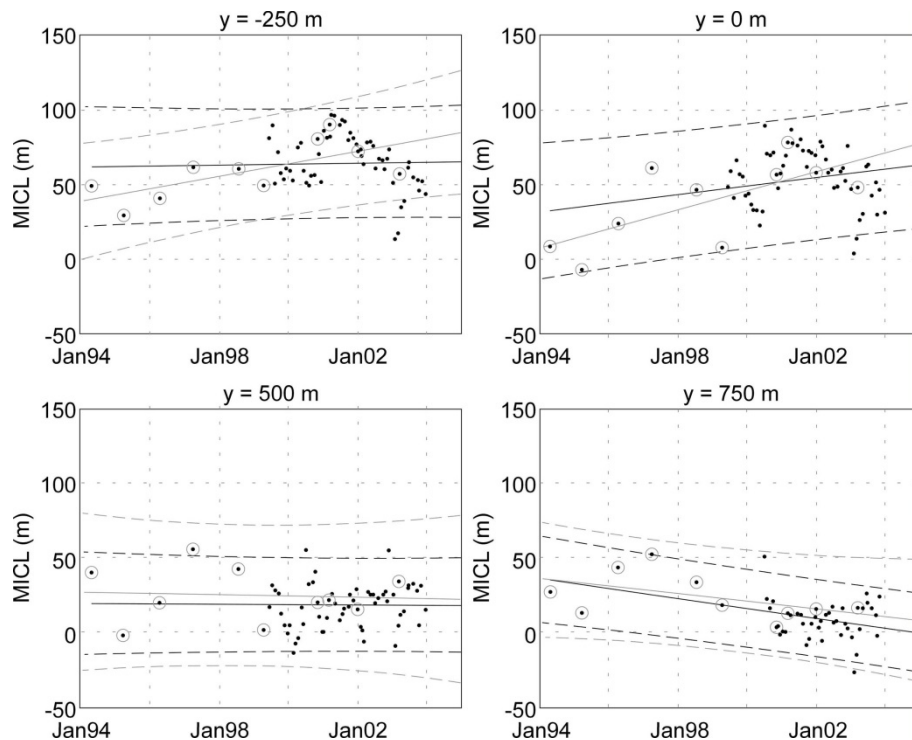


Fig. S3.6 Ten-year trend in Intertidal Momentary Coastline based on Jarkus data only (*solid grey line*) and on Jarkus and Argus data together (*solid black line*), at four locations alongshore (y). Circled dots represent annual traditional surveys (Jarkus), dots represent monthly video (Argus) measurements. Dashed lines represent the confidence intervals for both methods (Smit et al. 2007)

For studies covering the entire North Sea shore and focused on long-term change, merging datasets on coastal-state indicators requires upscaling to coastline lengths and classes that give meaningful visualizations for regional and transnational assessments, filtering out effects of inconsistencies. The end user must be made aware of limitations – uncertainty, error and indicative meaning.

The combination of derived and measured values and the assumption-based comparison of indicator proxies associated with a highly dynamic coastline introduce uncertainty and error; these may be relevant in process-response studies. Lateral and vertical uncertainties may be larger than short-term fluxes and mobility, but are generally inconsequential for overall comparisons in the context of climate change. For example, a large one-time error in legacy data on erosion may be small when translated to a value for trend/year in a long historical record (Leatherman 2003).

Shoreline mapping techniques applied to data sources are moving towards automation in association with technological advances. Mapping is also conducted more consistently, following protocols that minimize error from tidal and seasonal changes (Moore 2000; Leatherman 2003). Success of statistical analysis strongly depends on the total size of the data set characterized by the specific spatial coverage and resolution, temporal resolution and overall duration (Kroon et al. 2008).

Available datasets and some more detailed Netherlands and UK trends

At a national or regional level, the availability of datasets is a function of the severity and socio-economic impact of coastal erosion experienced. Hence there are no monitoring data for Norway, which is still rising as a result of isostatic rebound. In Sweden, the Swedish Geotechnical Institute (SGI) has coordinated monitoring of coastal erosion since 2002. A systematic assessment of coastal erosion (pocket beaches and bluffs of erodible sediment) including a complete set of maps at scale 1:250,000 was made by Rydell et al. (2004), using information supplied by coastal municipalities. In Denmark, the Danish Coastal Authority (Kystdirektoratet) and its predecessors have coordinated the monitoring since 1874. Its best available record is for the west coast of Jutland, which has been surveyed since 1957. Since 1969, profiles have been surveyed biannually just to the south at Skallingen. In Germany, coastal monitoring along the North Sea is

the responsibility of the coastal states Niedersachsen and Schleswig-Holstein; monitoring has been mainly project based, with most effort starting in the second half of the 20th century.

The Netherlands has an extensive subsiding and densely populated coastal lowlands. Along the western Netherlands coast, contemporary engineers accurately recorded continual erosion of Egmond aan Zee between 1686 and 1741, culminating in collapse of the church steeple during a major storm surge. The historical record shows many rows of houses disappearing, as a 100-m-wide strip of dunes eroded into the waves, despite coastal engineers' protective measures (Schoorl 1990, Fig. S3.7). [Before 1800, defences were developed largely autonomously and the coastline receded rapidly, 3 to 5 m/year, especially the northern and southern sections of the central coast.]

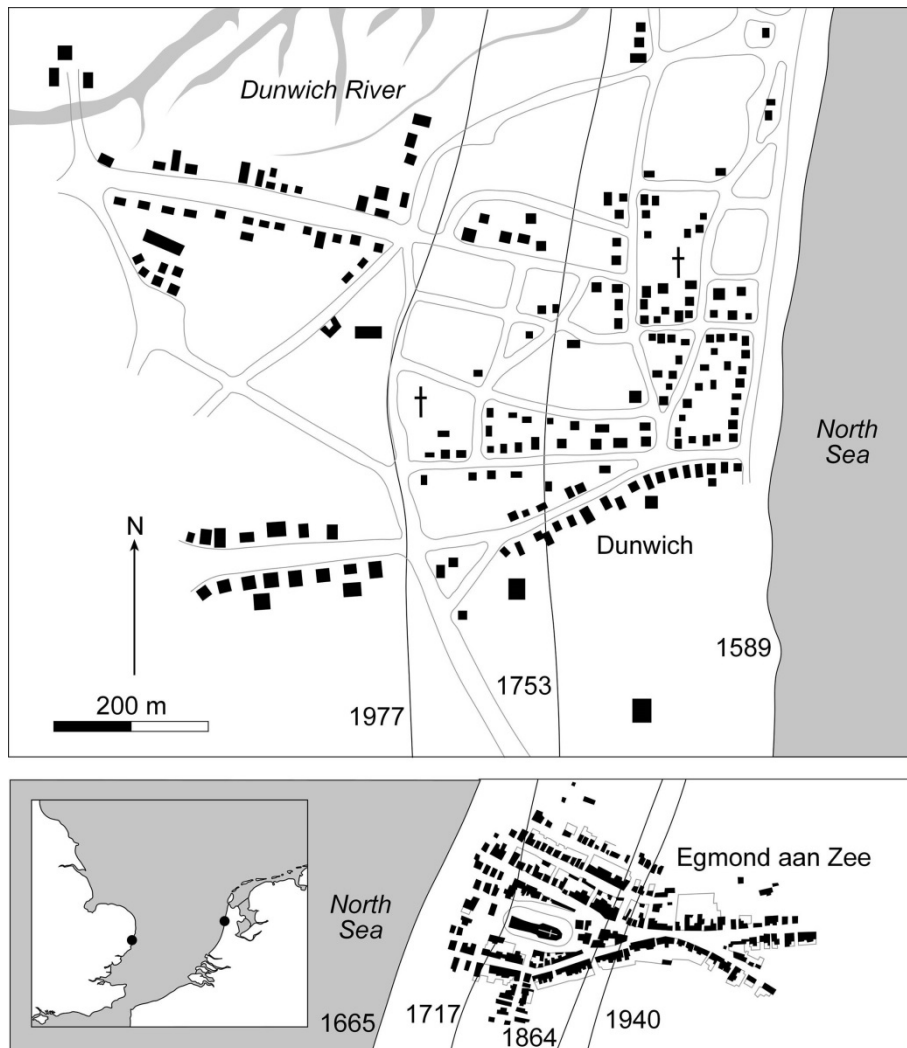


Fig. S3.7 Redrawn after Cooper and MacKenna 2008 (*upper*) and Schoorl 1990 (*lower*)

Annual coastline monitoring started in 1843, in the western Netherlands. It was instigated by Jan Blanken, inspector-general of the Department of Waterways and Public Works, formed half a century earlier by merging various local and governmental bodies. Between 1840 and 1857, 124 oak poles (Fig. S3.8) were driven about 3 m into the beach sand, initiating a network of beach poles with 1000-m spacing that spans the entire Dutch coast and still operates today. For each pole, cross-shore distances to the low- and high-water lines and dune foot have been logged annually ever since, creating a unique record of coastline change during a time that saw the transition from autonomous to increasingly anthropogenically-influenced processes and behaviour. Starting in 1965, some 1450 transects at 250-m spacing have been profiled near-annually from at least the frontal dune to about 1000 m seaward of the dune foot. Below low water, data come primarily from single-beam echo sounding. Above low-water, photogrammetry was used before 1996 and laser altimetry since.



Fig. S3.8 Beach pole

Netherlands monitoring is carried out under the auspices of Rijkswaterstaat, and all legacy and modern (JARKUS) data have been collated in OpenEarth by Deltares.

In Belgium, monitoring data (partly collected by universities of Ghent and Louvain) are held and analyzed by the Afdeling Kust of the Agentschap voor Maritieme Dienstverlening & Kust. In France, monitoring is project based and data have been collected primarily by researchers from the Université du Littoral Côte d'Opale in Dunkerque.

In Britain, maps from the late 16th century provide the oldest former positions of coastal bluffs and cliffs (Fig. S3.9). In England, the Environment Agency now coordinates biennial (winter and summer) cliff and beach profile monitoring at 1 km intervals between the Humber and Thames estuaries. This extends to much of the Channel coast as part of the Regional Coastal Monitoring Programme (<http://www.channelcoast.org/>). The Channel Coastal Observatory and British Geological Survey are also involved. Other monitoring includes photogrammetry, sub-seabed data from specific areas and recording coastal landslides to which sectors of the Channel coast are prone. Some LIDAR data have been obtained along the Scottish coast.

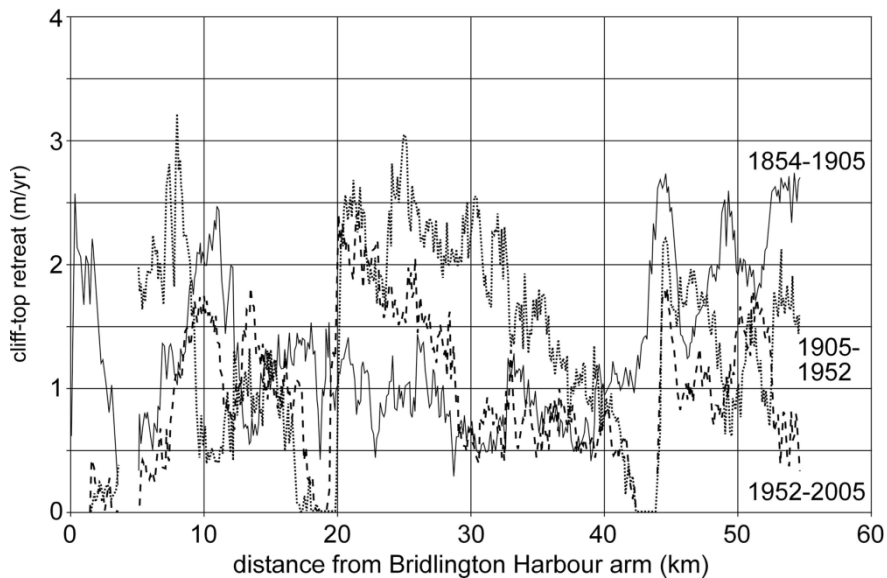


Fig. S3.9 Cliff-top retreat rates at Holderness calculated in three approximate-50-year periods. Areas defended in 2005 are in grey. Retreat rates vary within each zone. From Brown (2008)

The Holderness bluff coast (figure S3.9) eroded by more than 300 m in 150 years (Valentin 1954). At Happisburgh (figure S3.10) rates of erosion have increased to average ~ 8 m/y (Poulton et al. 2006), significantly faster than the long-term averages of 0.9 m/yr for North Norfolk (Cambers 1976; Thomalla and Vincent 2003), 2.3–3.5 m/yr for Benacre–Southwold and 0.9 m/yr for Dunwich–Minsmere (Brooks and Spencer 2010; figure S3.11 shows that coherent but large-scale shifts in erosional behavior can take place over alongshore distances of less than 1 km).

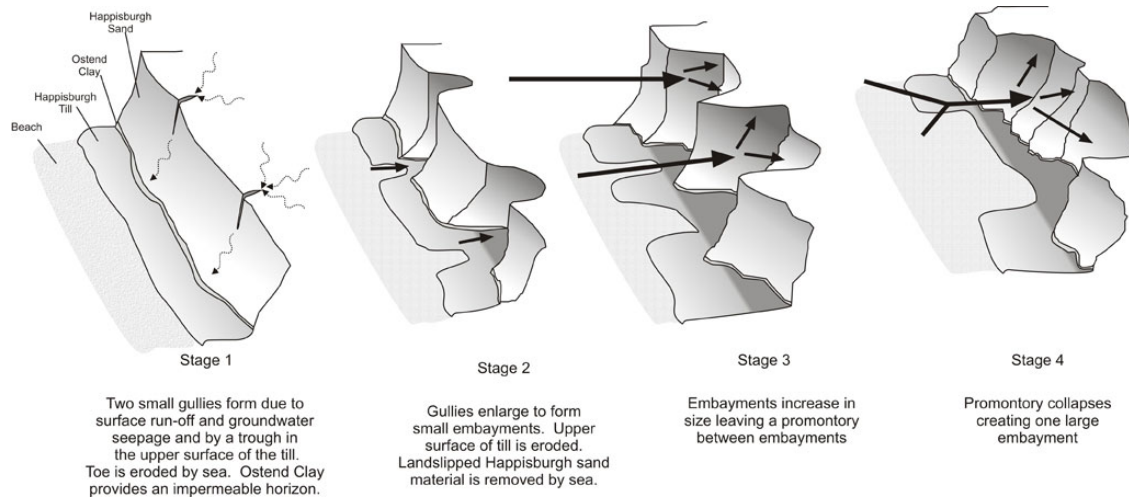


Fig. S3.10 Non-uniform bluff erosion, involving the cyclic formation of embayments that enlarge. This could involve block falls, mudflows and running sand (after Poulton et al. 2006 with the permission of the British Geological Survey)

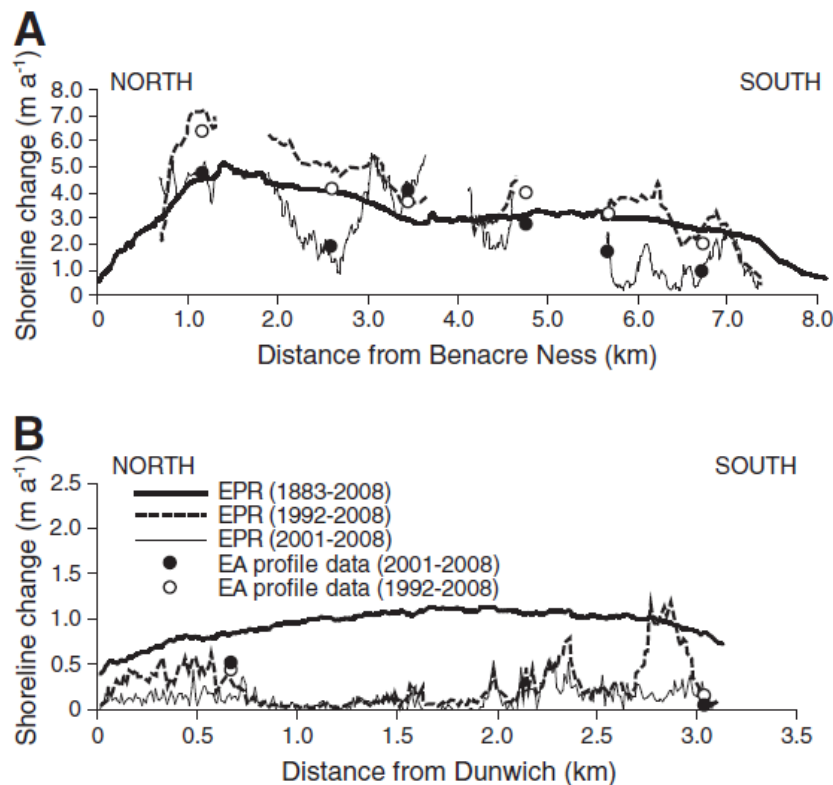


Fig. S3.11 Shoreline change (EPR m/yr) over three time periods using the Digital Shoreline Analysis System for A) Benacre to Southwold and B) Dunwich to Minsmere. Analysis is based upon digitized MHWS/cliff-base shorelines at 10 m alongshore intervals for 1883–2008. Analysis at 10 m intervals for cliff sections only, and based on top-of-cliff position from aerial photography. Also shown for 1992–2008 and 2001–2008 are rates of shoreline retreat at ~ 1 km spacing alongshore, from EA SDMS bi-annual profile surveys. Brooks and Spencer (2010)

Much of this information was collated in EUROSION, an EU project focusing on coastal erosion in Europe and including an assessment of the current status and trends of European coasts. This project, which was finalized in 2004, addressed four types of coastline: sandy or gravelly beaches, soft rocky coasts, hard rocky coasts, and muddy coasts. One of the main deliverables of EUROSION, a GIS layer showing coastal behavior at a scale of 1:100,000, has recently been updated and supplemented with data for Norway and Sweden in EMODnet-Geology, an ongoing EU project that brings together information on marine geology (Fig. S3.12).

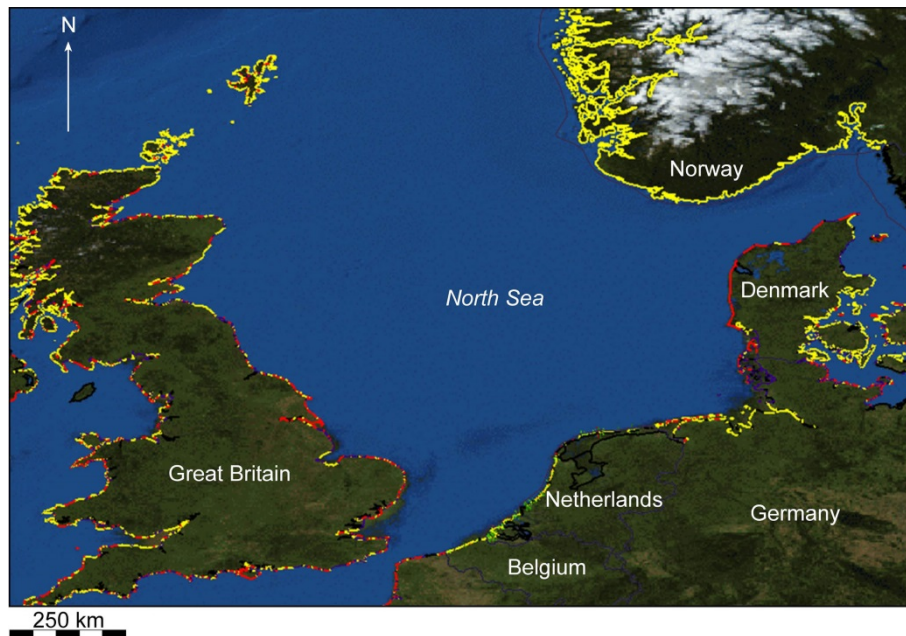


Fig. S3.12 EMODnet-Geology map of coastal erosion (*red*), stability (*yellow*) and accretion (*blue*) along the North Sea shore, updated from EUROSION

S3.4.1 Local studies of offshore morphology

Several studies concern dynamics of particular local offshore systems, supporting the main text Sect. 3.9.3 in asserting the role of tidal asymmetries with wave-enhanced transport, and that shoals evolve but long-term prediction is uncertain. In Texel inlet (Netherlands), modified by closure of a major part of its back-barrier basin, tidal currents redistribute sand; large gross transport rates but small morphological changes point to sediment recirculation. Sediment import into the basin results from net flood-dominated transport due to tidal asymmetry enhanced by landward-directed wind- and wave-driven flow (Elias et al. 2006). In the Deben estuary (Eastern England), ebb-tidal shoals exhibiting broadly cyclic behaviour on a 10-30-year timescale (Burningham and French, 2006). Short-term increases in total ebb-tidal delta volume are linked to annual variability in the northerly to northeasterly wind climate and consequent interaction between wave and tidal current-driven sediment transport. In the Thames estuary (Burningham and French 2011) individual central banks have shown lateral shifts and some of the outer banks show progressive elongation. Banks across the Suffolk shoreface show evidence of onshore and of offshore migration. At Vejers beach (Denmark) an outer bar loses sediment to longshore transport; an inner bar migrates offshore under the influence of waves and undertow (Aagaard et al. 2010). However, off Nordwijk (Netherlands) offshore bar migration is gradual and not coupled to individual storms (Walstra et al. 2013).

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