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Effect of Mesoscale Variability of Water Masses on Acoustic Wave Propagation in a Shallow Sea

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Abstract:

Anthropogenic noise in the sea is now classed as a pollutant alongside chemical pollution and marine litter in accordance with the Marine Strategy Framework Directive. Noise from shipping is a major contributor to the ambient noise levels in the ocean, particularly at low (<300Hz) frequencies. The properties of sound propagation in shallow waters are highly influenced by the marine physical environment. Ocean modelling plays an important role in underwater noise studies since it can provide high resolution water column parameters over large geographic areas. This study investigates the noise patterns and their temporal variations in the Celtic Sea by using a coupled ocean model (POLCOMS) and an acoustic model (HARCAM). A method to predict noise exposure experienced by marine animals is then developed, following an application for diving seals.

The ocean model is applied in the Celtic Sea to provide high-resolution 3D hourly temperature and salinity fields for the acoustic model. The model is validated against in-situ and satellite observations, giving high skills to simulate the water column structures. Sensitivity studies of modelled results to different atmospheric forcing are carried out in order to improve the accuracy of the model. The results show that the modelled sea surface temperature, stratification and water column structures are highly sensitive to the choice of surface forcing, especially in the summer time. The increase in resolution of surface forcing does not necessarily lead to more accurate results. The tidally frontal position is, however, insensitive to the forcing.

The variability of noise propagation is studied using the coupled model, demonstrating high dependence on oceanographic conditions, geographic location of sound source and its depth. In summer, when the source of sound is on the inshore side of the bottom front, the sound energy is mostly concentrated in the near-bottom layer. In winter, the sound from the same source is distributed more evenly in the vertical. When the source is on the seaward side of the front, the sound level from a shallow source is nearly uniform in the vertical and the transmission loss is significantly greater (~16dB at 40km distance) in summer than in winter. In contrast, sound energy from a deep source is trapped in the bottom cold water, leading to a much lower transmission loss (~20dB) in summer than in winter. Note that ~10dB fluctuation of sound energy is found during the deterioration of the thermocline in late autumn. Shallow sources (e.g. ships) are sensitive to the surface heat flux as it changes significantly the vertical temperature gradient, while tides play an important role in determining the TL
variability of deeper sources (e.g. pile driving) since they cause adjustments of positions of subsurface fronts.

The seasonal noise patterns radiated by a large cargo ship are modelled by relating the AIS ship track data and the coupled model, showing a clear influence of the seasonal thermocline and associated bottom fronts on shipping noise distribution. The noise propagates much further (tens of kilometres) in winter than in summer. The predicted shipping noise exposure perceived by grey seals shows strong step changes in the sound level during their descent/ascent through the water column. Since grey seals tend to be benthic foragers, a hypothesis that the step change in sound exposure may have negative impacts on their foraging behaviour is proposed for biological specialists.
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List of Abbreviations

AIS  Automatic Identification System
BADC  the British Atmospheric Data Centre
BoM  The bias of mean
Cor98  Towed undulating CTD Scanfish dataset
EOF  Empirical Orthogonal Function
EOF₁  Eigenvector of the Empirical Orthogonal Function
ERA40  ECMWF 40-year Reanalysis
ERA-interim  ECMWF Interim Reanalysis
ES2008  European shelf tide model
GOTM  General Ocean Turbulence Model
GUI  Graphical user interface
HARCAM  HODGSON and RAM composite acoustic model
JRA25  Japanese 25 year Re-Analysis project
JRA55  Japanese 55-year Reanalysis
KE  Kinetic energy
MCZs  Marine Conservation Zones
MPAs  Marine Protection Areas
NCEP  National Centers for Environmental Prediction
NEMO  Nucleus for European Modelling of the Ocean
NURC  NATO Research Centre
PDF  Probability density function
PE  Parabolic equation
PEA  Potential energy anomaly
POL  Proudman Oceanography Laboratory
POL3DB  Proudman Oceanographical Laboratory Three-Dimensional Baroclinic B-grid model
POLCOMS  Proudman Oceanographic Laboratory Coastal Ocean Modelling System
R  The correlation coefficient
RD  Receiver depth
RL  Received level
RMSe  The root mean square error
W  The Willmott skill parameter
SD  Source depth
SEL  Sound Exposure Level
ShipAIS  Shipping Automatic Identification System
SST  Sea surface temperature
TKE  Turbulence kinetic energy
TMD  Tide Model Driver
TL  Transmission loss
vPPM  Vertical piecewise parabolic method
Glossary

\( M_2 \)  principal lunar semidiurnal tidal constituent
\( S_2 \)  principal solar semidiurnal tidal constituent
\( h \)  water depth
\( \bar{u} \)  vertically averaged horizontal velocity
\( u \)  eastward velocity
\( v \)  northward velocity
\( \sigma \)  vertical level
\( \varphi \)  potential energy anomaly
\( g \)  gravitational acceleration
\( \rho \)  potential density
\( X_n \)  tidal velocities or elevations
\( H_n \)  average amplitude
\( f_n \)  nodal factor
\( \dot{\sigma} \)  angular speed (degree/hour)
\( g_n \)  local phase lag
\( V_n \)  tidal phase of 0 GMT on the start day of running
\( \theta_n \)  nodal angle
\( I \)  sound intensity
\( I_r \)  reference sound intensity
\( P \)  sound pressure
\( P_r \)  reference sound pressure
\( r \)  sound propagation range
\( \phi \)  potential function
\( \nabla^2 \)  Laplacian operator
\( c \)  sound speed
\( c_0 \)  reference sound speed
\( \omega \)  angular frequency of source
\( K \)  wavenumber
\( \delta \)  Dirac delta function
\( \lambda \)  wave length
\( k_0 \)  reference wavenumber
\( n \)  refraction index
\(F(x,y,z)\) pressure amplitude function
\(G(x,y,z)\) phase function
\(M_n\) model field
\(O_n\) observational field
\(\bar{M}\) mean of model field
\(\bar{O}\) mean of observational field
\(\sigma_m\) standard deviation of model field
\(\sigma_o\) standard deviation of observational field
\(W\) Willmott skill parameter
\(q_{\text{in}}\) inward hear flux
\(q_s\) solar radiation available to the surface
\(c_c\) cloud cover coefficient
\(\alpha\) sea surface albedo
\(\Sigma\) covariance matrix of EOF
\(\lambda_i\) eigenvalue
\(\hat{a}_i\) principle component
\(L_s\) the narrow band spectrum of source
\(l\) ship’s length in feet
\(L_{s0}\) reference spectrum defining an average ship as one with a speed of 12knot and a length of 300 feet
\(v_s\) ship’s speed in knots
\(P_{RL}\) received sound pressure
\(R_i\) received mean square pressure
\(T_i\) time interval in seconds
\(q_t\) outgoing heat flux at sea surface
\(q_{\text{in}}\) incoming heat flux at sea surface
\(\tau_i\) wind stress
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Author’s Declaration

At no time during the registration for the degree of Doctor of Philosophy has the author been registered for any other University award. Work submitted for this research degree at the Plymouth University has not formed part of any other degree either at Plymouth University or at another establishment.

Relevant seminars and conferences were attended at which work was often presented. Several papers were prepared and submitted to the refereed journals, which are at different status at the time of completion of the thesis. In joint publications the author has made a distinct and substantial contribution.

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Chapter 1

1 Introduction

1.1 Background

Changes in the oceanic acoustic environment arising from anthropogenic activities (e.g. sonar systems, seismic exploration, increasing maritime traffic and offshore renewable energy developments) have received significant attention in recent times. The regulatory bodies have placed additional restrictions on uses of underwater sound, especially in shallow shelf seas (European Commission Decision, 01/09/2010). Underwater noise is now classed as a pollutant alongside chemical pollution and marine litter in accordance with the Marine Strategy Framework Directive (MSFD).

Underwater acoustics is used increasingly in oceanographic and environmental studies, and continues to play a crucial role in defence. It is also widely used for communication, navigation and identification of objects both by humans and marine mammals (Katsnelson et al., 2012) and for investigating detrimental effects of anthropogenic activities (e.g. pile driving, seismic survey and ships) on marine animals (Richardson et al., 1995; Southall et al., 2007). It encounters challenges in shallow waters, however, due to high dependence on oceanographic conditions and complicated boundary interactions with the sea surface and seabed (Urick, 1983; Jensen et al., 2011). Since underwater acoustic propagation modelling has been developed intensively for many decades, the biggest difficulty is not related to the modelling tools themselves, but rather to the lack of detailed environmental parameters as
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inputs of acoustic models (Jensen et al., 2011). To a large extent, the accuracy of shallow water acoustic modelling is determined by the quality of the water column parameters and sediment data.

The patterns and parameters of acoustic propagation, as well as the level of transmission loss (hereafter TL) in shallow waters, are highly sensitive to the inhomogeneities of temperature and salinity (Katsnelson et al., 2012). Due to their dynamic nature, the ocean features typical for the shelf sea, such as fronts, eddies, filaments, variations in the seasonal thermocline, internal waves and similar highly variable features, generate a highly variable environment for acoustic wave propagation. There is mounting evidence that these oceanic features can cause fluctuations of sound energy to a variety of degrees (e.g. Lysanov et al., 1989; Heathershaw et al., 1990; Lynch et al., 2003; Xu et al., 2009; Lermusiaux et al., 2010). These ocean processes occur in various spatial and temporal scales, which create great challenges for acoustic modelling in shallow coastal regions. Understanding the mechanism of sound propagation in shallow seas under various climatic and weather conditions is, now, of particular concern since commercial and civilian activities in coastal waters have been increasing rapidly. The fluctuations in the level of sound energy resulted from oceanic variations can be transferred into noise fields, hence leading to different noise patterns in time and space. A focus proposed by MSFD is to understand how environmental conditions affect noise patterns and their scales in time and space since little is known about it.

Acoustic modelling requires an accurate representation of water column data through either observation or modelling. Due to constraints of field experiments for providing adequate temporal and spatial coverage of environmental data, coupled ocean-acoustic models play an important role in the application of acoustic modelling. Ocean models combined with data assimilation methods are often used to provide water column parameters for acoustic models
(e.g. Heathershaw et al., 1990; Lam et al., 2009; Rixen et al., 2009; Xu et al., 2009 and Lermusiaux et al., 1999, 2006, 2010). Such coupled modelling systems offer capabilities of studying underwater sound propagation over various spatio-temporal scales.

Many marine animals are very sensitive to sound as they produce and perceive sound for a number of activities (e.g. communication and predation). Much evidence has been found to suggest that elevated noise levels have great negative effects on marine animals (e.g. Parks et al., 2007; Wright, 2008; Jensen et al., 2009; Rolland et al., 2012). Knowledge of understanding the characteristics of noise and its temporal and spatial variability is, thus, of particular importance to the protection of marine life.

![Fig. 1.1 Map of study area (red rectangle) showing the bathymetry](image)

The geographic area selected for this study is the Celtic Sea (see Fig. 1.1), situated in the southwest of the British Isles with a shape of semi-enclosed embayment. It features shallow bathymetry, strong seasonality of heat content, sea surface temperature (SST) and stratification, intense bottom fronts and density-driven currents in summer (Pingree, 1980;
Brown et al., 2003). It also has importance as a shipping thoroughfare, intensive shipping activities result from the sea being a transition zone from Atlantic waters to the shallow coastal waters of the European Continental Shelf. Furthermore, the sea is located in one of four Marine Conservation Zones (MCZs) in UK waters area and is of great ecological importance to the seals, harbour porpoises, common dolphins, bottlenose dolphins and minke whales since abundant species have been observed in the region (Reid et al., 2003; Hammond et al., 2008).

In terms of oceanography, the Celtic Sea has been studied intensively based on both numerical modelling and observations for several decades and the underlying physics have been well established. Many modelling systems have been also developed and used by different research groups and organisations. Different regional models (e.g. Holt and Umlauf, 2008; Shapiro, 2011; Maraldi et al., 2013) covering the Celtic Sea are also implemented to investigate the physical processes, provided with error qualifications. The development of the ocean models has being always of particular interest in improving continuously the accuracy and predictability of the models, which is dependent on a number of factors. Although a standard model has been implemented in the Celtic Sea by Shapiro (2011), the model simulations have not been validated yet. This is complemented by this study where the model outputs are compared with intense observational data. In addition, the model results are further improved by implementing new tidal forcing. Compared with previous studies (e.g. Holt et al., 2005; O’Neill et al., 2012) a better surface forcing which increases the degree of the model accuracy is found by performing sensitivity studies.

With regards to the underwater sound propagation in general, modelling tools are robust and well developed. The variability of sound propagation medium, which depends strongly on the geographical location, plays more important role in determining behaviours of sound
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propagation. Many studies of investigating the effects of oceanic process on sound propagation have been carried out in different shelf seas worldwide, but excluding the Celtic Sea. This project fills the gap and examines the variability of sound propagation resulted from changes of water column structure in the Celtic Sea.

Anthropogenic noise is now a hot topic and sound mapping is a major step to monitor noise pollution in shelf seas as stated by MSFD. Development of prediction tools to map sound levels in oceans is on the way. A mapping tool which uses shipping AIS data and sound propagation models with simplified environmental inputs has been developed by Erbe et al. (2012) to monitor noise intensity. This project uses high resolution coupled ocean-acoustic modelling system and investigates spatial changes of shipping noise by predicting the noise patterns in different seasons, which contributes to the understanding of how the oceanic conditions affect the noise distribution in the Celtic Sea. The sound maps, which are very useful for spatial planning, can be used to identify high noise areas where anthropogenic activities overlap intensely with vulnerable marine ecosystem.

The impact of noise on behaviours of animals is a particular concern as proposed by MSFD, however little is known about it. Studies of behavioural disturbance to animals require constant noise exposure experienced by animals. Previous researches (e.g. Castellote et al., 2012; Wale et al., 2013) are focused primarily on controlled experiments, which have several limits (e.g. a few species, short time durations and expensive). A new method, which combines a coupled ocean-acoustic modelling system with shipping Automatic Identification System (AIS) data and seal diving data, has been developed in this study, in order to predict noise patterns of animals along their travelling path. Strong step changes in sound level are found when seals dive through the water column, which might have effects on their foraging behaviours. This method, which links noise intensity with animal diving behaviours, can be
used to reconstruct historical noise patterns of animals as long as the AIS data and seal diving data are provided.

1.2 Aims and objectives

This work is divided into three main sections, taking the route of 1) ocean modelling, 2) coupled ocean-acoustic modelling and 3) an application of predicting shipping noise exposure experienced by animals. The aims of each respective section of this study are as follows:

- **Ocean modelling:** To improve the accuracy of the ocean model in simulating the water column structures in the Celtic Sea and to provide high quality environmental data for the acoustic model by using the improved ocean model.

- **Coupled ocean-acoustic modelling:** To examine how the oceanic processes in the Celtic Sea affect underwater sound propagation.

- **Application:** To understand the potential seasonal variability of shipping noise patterns in the Celtic Sea and the potential sound exposure experienced by seals.

In order to achieve the aims outlined above, the following objectives will be addressed:

- A standard version of the ocean model for the Eastern Celtic Sea are extended to a larger area and the tidal and full-model predictions are validated against observational data.

- New tidal calculations of the ocean model are implemented and validated to improve the tidal prediction of the model.

- Sensitivity of model outputs to different meteorological forcing obtained from different sources is examined to obtain optimised surface forcing for the model.

- Sound transmission losses under different oceanographic conditions and acoustic geometric configurations are predicted to examine the effects of the oceanic features on sound propagation in the Celtic Sea.
• The three-dimensional shipping noise patterns by a large commercial cargo ship in different seasons are simulated.

• Historical noise exposure in a shorter range experienced by seals along their diving profiles is reconstructed.

1.3 Thesis outlines

Chapter 2 continues with the literature review, covering the physics in the Celtic Sea, ocean modellling, coupled ocean-acoustic modelling, shipping noise and effects on marine animals.

Chapter 3 describes the main methodology used by this work, including the physics of the oceanographic model POLCOMS and the acoustic model HARCAM, details of model set-up and statistical analysis for data processing.

Chapter 4 contains the results of ocean modelling both in barotropic (tidal) and fully 3D model with 2km horizontal resolution. The model is rigorously validated against observations. A significant effort is dedicated to the sensitivity studies of model results to different meteorological forcing.

Chapter 5 contains the main results of coupled ocean-acoustic modelling. The variability of propagation loss resulting from oceanic features are analysed from seasonal down to hourly scales.

Chapter 6 applies the coupled model for investigating the spatio-temporal pattern of shipping noise, simulates the noise intensity at various locations and discusses how this might affect marine mammals, such as seals.

Chapter 7 gives the overall summary of this project including the results, discussions, conclusions and the future research.
Chapter 2

2 Literature review

2.1 Physics in the Celtic Sea

The geographic area selected for this study is the Eastern Celtic Sea shown in Fig. 1.1. It is encompassed by the coasts of southern Ireland, south-west Wales, Cornwall, and the St George’s Channel, the English Channel and the 200m isobath of the continental shelf linking with the North Atlantic (Thompson & Pugh, 1986). The bathymetry in this area (red rectangular box in Fig. 1.1) is shallow with depth varying from a few meters around coasts to more than 120m in the centre of Celtic Deep.

The Celtic Sea is a tidally dominated water basin, with the tidal stream amplitude varying from 0.25ms$^{-1}$ in the southwest to 2ms$^{-1}$ in the Bristol Channel (Huntley, 1980; Manning et al., 2010). As described by Doodson and Corkan (1932) the dominant semi-diurnal tidal constituents $M_2$ and $S_2$ represent the main tidal features in the region of the Celtic Sea. At spring tides the overlapping of $M_2$ and $S_2$ in phase leads to tides with an amplitude 1.33 times the value of $M_2$, while it is 0.67 times the value of $M_2$ at neap tides (Pingree, 1980). Tides, to a large extent, provide the most energetic processes in determining the seasonal features in the Celtic Sea (e.g. seasonal thermal stratifications and tidal fronts).

Internal waves are of notable importance to the effect of the circulation patterns, an important influence on the physical structure of the Celtic Sea (Holt and Thorpe, 1997) and strong effects on acoustic propagation (Katsnelson et al., 2012). It has been also known that the
south-west border of the Celtic sea at the 200m isobath is the region with significant internal
tide activities (Pingree et al., 1983). In this region, the tidal currents are remarkably strong
due to the extremely rough topography, propagating with the components along and cross the
slope. According to the observations in 1983, Pingree and Mardell (1985) reported that peak
to trough amplitudes of waves were noted to exceed 50m with periods of 15 min to 1 hour,
speeds of 0.7ms\(^{-1}\) and inter packet separations of 30km. A study from Pingree and New (1995)
using remote sensing has shown that the internal tide energy propagates coherently up to
many hundreds of kilometres across the Celtic Sea. Direct observations along a transect of
300km by Inall et al. (2011) have confirmed that the internal tide across the broad continental
shelf of the Celtic Sea is coherent over more than 170km, with many wavelengths. Note that
in their study the shoreward energy decay scale is estimated as 42km. The wavelength-
averaged energy decay rate near the shelf is estimated to be \(2.08 \times 10^{-7}\) W\(Kg\)\(^{-1}\) based on the
measured hydrographic and velocity data, in close agreement with the tidally and vertically
averaged measurements in this region.

The sea is strongly stratified with a sharp thermocline from April to November (Pingree,
1980), which adds to the formation of density-driven currents. According to observations
conducted by Horsburgh et al. (1998), a strong baroclinic jet-like circulation was examined in
the St George’s Channel associated with the margins of a cold pool with dense bottom water.
Following Brown et al. (2003), it has been further confirmed that the summer circulation in
the Celtic Sea is dominated by an intense and persistent cyclonic baroclinic jet-like flow. At
the margin of the cold pool where the bottom front is located, a sharp density gradient
between the cold pool and surrounding waters is formed and results in a baroclinic jet-like
flow (~30cms\(^{-1}\)), which circulates around the cold pool and extends 170 km from the Irish
coast as far as to the Scilly Isles (Brown et al., 2003; Young et al., 2004). A seasonal
baroclinic circulation travelling from the south of the Celtic Sea to the St George’s Channel
along the northern coast of Cornwall has been also identified. The westward flow across the St George’s Channel extends to the Celtic Deep southwards into the Celtic Sea directly, and west along the Irish coast as the pattern of intrusion (Brown et al., 2003).

Seasonal stratification is a significant feature of the European shelf seas, including the Celtic Sea. It develops in early summer as the surface solar heating increases and the turbulent kinetic energy of the waters decreases. Eventually it is broken down in the later autumn by the overturning and increasing wind speed (Simpson and Hunter, 1974; Elliott et al., 1991). Thermal stratification occurs where there is insufficient mechanical energy due to wind near the surface and due to bottom tidal mixing against the surface buoyancy produced by solar heating (Fernhead, 1975). Initially, the investigation of shelf sea stratification came from the energy consideration. A critical value $\log_{10} \left( \frac{h}{\bar{u}^2} \right) = 2.7 \pm 0.4$ for the European Continental Shelf of stratifications subject to water depth $h$ and vertically averaged horizontal velocity $\bar{u}$ was given by Simpson and Hunter (1974). A result lower than the critical value refers to well mixed conditions, while a higher than critical number is associated with stratified waters. Subsequently it has been applied successfully in the European Shelf Sea by Bowers and Simpson (1987) and the mean position of stratification has been also well established for the European shelf seas. However, this simple energetic theory overestimates substantially the variability in frontal positions occurring from the spring–neap cycle, which results both from using a fixed mixing efficiency and from the lack of a time evolution of the buoyancy (Holt and Umlauf, 2008). The thermocline appears first in the region of weak tidal mixing near Nymphe Bank (close to the southeast coastlines of Ireland) and extends eastwards to the Bristol Channel while the development is relatively delayed near the north Cornish coast due to increased tidal mixing in the coastal area (Pingree, 1980). Conversely, the pattern of thermocline erosion is a reverse of the development. The difference between the development
and retreat of the thermocline is that the development is more rapid occurring in a few weeks, whereas the erosion is comparatively longer, taking three months to cross the region.

The potential energy anomaly $\varphi$ (Simpson and Bowers, 1981) is a useful measure of stratification and frontal positions, which is defined by equation (2.1):

$$\varphi = \frac{g}{h} \int_{z=-h}^{0} z(\rho(T,S) - \rho(\bar{T},\bar{S}))dz$$

where $g$ is the gravitational acceleration, $\rho$ is the density, $h$ is the water depth and the overbar represents the depth mean. The potential energy anomaly is a vertically integrated measure of energy required to completely mix the water column. $\varphi<0$ indicates well mixed water while $\varphi>0$ represents the stratified region. $\varphi=0$ can be used, therefore, to mark the position of fronts. The magnitude of PEA is also a measure of the strength of stratification with many applications, Simpson and Bowers (1981), Holt and Proctor (2008) for example.

The sediments in the Celtic Sea are dominated by sand bottom type (Manning et al., 2010) and their distribution is shown in Fig. 2.1 (reproduced from Duggan, 2010). The sediments are composed of approximately 10 bottom types, the most common of which are gravelly sand, sand and sandy gravel. The gravelly sand and sandy gravel are most concentrated close to the northern coastlines of Cornwall and regions south to the Celtic Deep whilst the sand and muddy sand are found frequently in deeper waters (e.g. the Celtic Deep).
2.2 Ocean modelling

2.2.1 Numerical modelling in the Celtic Sea

The ocean model used for this study is the Proudman Oceanographic Laboratory Coastal Ocean Modelling System (POLCOMS) developed by Holt and James (2001). It is a three-dimensional primitive equation finite difference model that has been used successfully for modelling different regions of the world ocean, such as the European continental shelf (e.g. Holt and James, 2001; Holt and Umlauf, 2008; Holt et al., 2010), the Black Sea (Enriquez, et al., 2005) and the Celtic Sea (Shapiro, 2011). POLCOMS has been also used operationally by the UK Met Office for the European shelf seas (Bell, 2012). Although it is being largely superseded by NEMO (Nucleus for European Modelling of the Ocean) for the operational usage, POLCOMS is still an advanced model used widely for other applications. For instance,
it is coupled with biological models to investigate the potential effects of climate change on ecological systems (e.g. Holt et al., 2014).

The Celtic Sea has been intensively studied for several decades using numerical modelling. Pingree and Maddock (1977) were the first to develop a numerical model of tides in this area and reproduced the primary tidal features of the Celtic Sea. A three-dimensional model developed by Davies and Jones (1992) using various parameterisations of eddy viscosity reproduced successfully the spatial distribution of tides covering the Celtic Sea. This model can simulate accurately the elevation over the entire domain. The Proudman Oceanographical Laboratory Three-Dimensional Baroclinic B-grid model (POL3DB) developed by Holt and James (2001) demonstrated an improving capability in modelling the baroclinic processes in the northwest of European continental shelf, including the Celtic Sea. The seasonal cycle of temperature and salinity were well reproduced. It can also predict the residual currents reasonably. Subsequently the model errors of modelling the hydrodynamics in the Northwest European continental shelf were quantified by Holt et al. (2005), giving high predictive skills in simulating the tidal flow, SST, water column temperature and etc. This model was then further improved by adding the Laplacian diffusion term (with Smagorinsky (1963) algorithm) to the force term of the momentum equation and scalar transport equation over which eddies were resolved more accurately (Holt et al., 2006), and by coupling the General Ocean Turbulence Model (GOTM; Umlauf and Burchard, 2005) through which the accuracy of modelling the tidal mixing fronts and seasonal stratification was promoted (Holt and Umlauf, 2008). Recently, a 3D regional POLCOMS model has been applied successfully in the Celtic Sea by Shapiro (2011) to predict the circulation pattern. POLCOMS has been also coupled with biological models to investigate the impacts of physical processes on ecosystem (e.g. Holt et al., 2014).
2.2.2 Sensitivity Study

In the context of computational-based simulation, the quality of ocean simulations depends on a number of factors such as approximations in governing equations, errors introduced by the numerical scheme, uncertainties in input parameters, and atmospheric forcing. Relations between those uncertainties and the accuracy of a model’s output have been a topic for a number of studies. Wright and Stocker (1992) examined sensitivity of the meridional fluxes of heat and water as functions of model parameters as well as the significance of various model simplifications. They found that the results are relatively insensitive to the value of the horizontal diffusion coefficient provided it is of the order of $10^3$ m$^2$/s or smaller. Incorporation of the realistic wind stress improves significantly the comparison with observational estimates. The effects of the vertical discretisation schemes on the accuracy of density driven currents was studied by Ezer and Mellor (2004), Legg et al. (2006) and Shapiro et al. (2013). The effect of variation in initial and open boundary conditions on circulation in the North East Atlantic was studied by Wakelin et al (2009). They found that on the continental shelf, the effect of using different ocean model initial and boundary conditions was small, whilst it was significant in the deeper oceanic regions. Berntsen et al. (2010) studied the sensitivity of non-hydrostatic effects to the grid size and found that non-hydrostatic effects are small for horizontal grid sizes larger than 50m.

Atmospheric forcing plays a significant role in controlling the dynamics of shelf seas, in the form of surface fluxes of heat, momentum, precipitation and evaporation. Sensitivity studies are becoming increasingly important due to their particular role in improving the accuracy of shelf sea models. There have been a large number of sensitivity studies of model simulations to atmospheric forcing either from different sources (e.g. Cravatte and Menkes, 2009; Shapiro et al., 2012; Chen et al., 2013) or from the same source with dissimilar spatiotemporal resolutions (e.g. Brossier et al., 2011, 2012; O’Neill et al., 2012), concluding
that the hydrodynamic and thermodynamic responses of seas are highly sensitive to the choice of surface forcing.

It is likely that the strongest effect on the quality of modelling of currents in a non-tidal sea is the accurate representation of the meteorological parameters. For example, the basin-wide circulation pattern and the temperature structure in the Black Sea produced by the same model are highly dependent on the source of the meteorological input (Shapiro, 2011). Atmospheric data of approximately the same resolution from two different sources, National Centers for Environmental Prediction (NCEP) and Japanese 25 year Re-Analysis project (JRA25) sometimes provides very different circulation patterns and water column stratification.

In a tidally active sea, Young and Holt (2007) studied the sensitivity of salinity fields produced by a 3D numerical model to boundary conditions and precipitation/evaporation in the Irish Sea. They concluded that investigating the sensitivity to the details of the meteorological forcing would be an obvious next step. A recent sensitivity study by O’Neill et al. (2012) examined the effect of varying ocean model resolution on the model skill in representing sea surface temperature and salinity in the Irish Sea. The authors showed that the NEMO model with 7 km resolution performed very similarly to the 1.8 km POLCOMS model when both used the same atmospheric forcing data set. The representation of the sea surface temperature was improved when meteorological forcing with higher spatial and temporal resolution was used. In particular, the use of higher resolution forcing reduced the RMS error in sea surface temperature by 20-30%.
2.3 Acoustic modelling and underwater sound propagation

The properties of underwater sound propagation are highly influenced by the marine physical environment. The pathways and the TL of acoustic energy are determined by the three dimensional distribution of sound speed, the geoacoustic properties of the seabed and sea surface roughness (Hamilton et al., 1982; Jensen et al., 2011 and others). The direction and intensity of sound propagation is determined by the sound speed gradients in the water column, which in turn are dependent upon variations in temperature, salinity and pressure. The shallow seas are an extremely dynamic environment featuring complex bottom topography, strong density and thermal fronts, eddies, filaments and other mesoscale features, which can cause energy fluctuations of sound to various degrees. Physical processes in the shelf seas which modify the temperature and salinity (and hence sound speed) distribution range from the mesoscale (days and weeks) to seasonal (months) to climatic scales (tens of years) in time and from hundreds of metres to tens of kilometres in space (Huang, 2009). Ocean features such as lenses of warm and saline waters observed in the North Atlantic disturb the location of acoustic shadow and silence zones by as much as 20-30km (Lysanov et al., 1989).

Sound energy loss and absorption occur when sound signals propagate from water medium to boundaries (e.g. sea surface and sediments). Compared with deep water, interactions of sound with the sea surface and seabed in shallow waters are increased significantly due to much shorter propagation distance in the vertical direction of the water column. This requires detailed boundary information for modelling sound propagation, particularly the seabed information over large spatial scales which is not generally available. The geoacoustic information of seabed used for sound propagation modelling must contain the sediment parameters (e.g. densities and sound speeds) through the effective depth, depending on the
frequency of sound. At high frequencies details of the bottom are required through the upper few metres as sound energy can be absorbed rapidly whereas at low frequencies information must be provided over the whole bottom and even the underlying rocks (Jensen et al., 2011). Work to explore completely the geoacoustic parameters in large scales is extremely expensive and impractical. A single bottom type (range-independent) was, therefore, adopted by many previous studies (e.g. Xu et al., 2009; Lam et al., 2009) due to a lack of sediment parameters. However, a study from Lermusiaux et al. (2010) has revealed that the TL modelled by using hybrid bottom types shows much better agreement with field measurements compared with any single bottom simulation. Consequently, a better representation of the sediments either by modelling or through measurements is indispensable to improve the prediction of sound propagation in shallow waters.

Underwater acoustic propagation modelling and simulation has been developed for many decades. Although it is a mature technique in deep oceans, the transformation from deep water to shallow seas encounters great challenges due to complicated variations of environmental conditions (Katsnelson, et al., 2012). A large number of propagation models exist, each of which has its own applicability and limits. Based on the underlying physical theory, these models are categorised as ray theory models, normal mode models and those using wavenumber integration techniques and parabolic equations. An extensive review of existing models, including the model theory, validation, advantages and disadvantages, can be found in Etter (2001).

Because of the rapidly increasing number of modelled and observed ocean data, the super-ensemble techniques which combine different model results along with the data assimilation have been used widely to reduce the uncertainties of coupled ocean-acoustic models, hence increasing the predictability (e.g. Lermusiaux, et al., 1999, 2006; Rixen, et al., 2009). A
A modelling study of sound propagation through an ocean front was carried out in the early work by Heathershaw et al. (1991). In this study a numerical ocean model was set up in an idealized ocean domain to provide environmental data for input to a range-dependent acoustic model. It was found that dependent upon sound source/receiver depth combinations, the effect of the front and eddies was to increase propagation loss by as much as 10 – 20dB. Using a coupled ocean-acoustic model, a fifteen days simulation was carried out by Xu et al. (2009) to investigate the effects of wind and tides on sound propagation. The research revealed that wind can cause an increase of transmission loss fluctuation (~20dB) at larger frequencies while tides introduce bigger energy fluctuation (~20dB) at lower frequencies, but the significant energy fluctuations are confined only to some specific points (e.g. the null points). The coupled ocean-acoustic modelling was also used to quantify the uncertainties of such a coupled system (Lermusiaux, et al., 2010) and the real-time application of propagation prediction at sea for operational purpose (Lam, et al., 2009).

A number of field experiments have been conducted in different regions of the ocean in order to assess how environmental factors impact acoustic propagation. The acoustic experiments carried out in Korean coastal regions by Abbot et al. (2001) demonstrated that variations in TL can be as high as 20-40dB due to changes in the type of bottom sediments, 10-15dB due to variations in the source depth and 8dB due to different frontal structures. Fine-scale features (e.g. internal tides and waves) can also result in significant fluctuations of sound energy of the order of 5-20dB (Lynch et al., 2006). Note that the horizontal refraction angle exceeded 1° in the experiments by Weinberg and Clark (1980). The study at the New England shelf break front (Lynch et al., 2003) based on the measurement of environmental data (temperature and salinity) and modelling of acoustic propagation, showed that diurnal and seasonal variations in the oceanic environment have strong effects on the TL. In the East China Sea shallow water area, a standard deviation of about 2dB of the TL fields was
observed (Abbot et al., 2003) while it was ~4-5dB in the central East China Sea, resulting from the fine-scale variations of a warm filament (Ramp, et al., 2004).

Internal wave is of great importance on underwater sound propagation as it creates fluctuations of sound energy when sound waves cross it, in the mechanisms of changing the propagation path and encountering scattering (Katsnelson et al., 2012). Experiments conducted by Rouseff et al. (2008) have shown that a new acoustic path is generated above the acoustic source when the internal wave passes. This indicates that internal waves change the propagation path of the sound waves as they introduce step changes in sound speed at the interface where strong reflection occurs. Acoustic simulation results by Srideve et al. (2010) have revealed that sound intensity loss of eigenrays varies from 2.86dB to 15.59dB in the presence of internal waves in the Northern Indian Ocean. In their study the maximum of loss (38.48dB) is observed at the bottom due to strong bottom interactions. Scales of internal waves vary greatly in time and space, which creates difficulties to determine the extent in changes of sound intensity. In addition, in regions near coasts and strong generation areas internals waves are able to develop into sharp fronts and thermocline, resulting in thermocline being shoaled significantly by tens of meters in only a few minutes (Katsnelson et al., 2012).

In many cases, internal waves, treated as random inhomogeneities, cause sound intensity biases and fluctuations by orders of 5-20dB (Lynch et al., 2006).

The largest moribund sandbanks on the earth exist in the Celtic Sea, consisting of a number of linear tidal sand ridges. The linear tidal sand ridges field covers an area of 65,000km$^2$, with a southeast-northwest trend in axes extending from the -200m isobaths to the southwest of the Isles of Scilly. The largest ridge in this region reaches 200km in length, 55m in height and 15km in width (Scourse et al., 2009). Such large sandbanks are made primarily of sand sediments. Gravel occurs between the linear tidal sand ridges, with mud being accumulated...
on the top of the gravel. Such mixture in the bottom type has strong impact on sound propagation. In a recent sensitivity study by Lermusiaux et al. (2010) it is shown that uncertainties in sediment sound-speed have a larger effect on TL than that in sediment attenuations. Diverse thicknesses of sediments lead to only limited effects on the TL. However, it also shows that stronger perturbations of sediment parameters (such as between clay and muddy-sand) could result in greater variations of TL, of about 5dB by comparing to observations.

The acoustic model used for this study is the HODGSON and RAM composite acoustic model (HARCAM), which has been validated formally by the U.K. Ministry of Defence over a variety of frequencies (10Hz - 500kHz) in both shallow and deep waters (Etter, 2001). It is a software engine for the Naval Tactical Decision Aid WADER-32 utilised operationally by the Royal Navy. HARCAM is an unclassified software package available as a Commercial Off The Shelf solution which allows users to calculate the acoustic signal propagation loss and the level of sound exposure at a distance from the source. RAM is a robust parabolic equation model developed by Mr Michael J Collins while HODGSON is a ray-based range dependent model. The two models form a symbiotic relationship in order to produce accurate TL efficiently over large frequency bands. The model also calculates the sea surface losses, the absorption losses, the bottom attenuation and loss and the reverberation.

2.4 Shipping noise and marine organisms

Noise from shipping is a major contributor to the ambient noise levels in the ocean, particularly at low (<300Hz) frequencies (Richardson et al., 1995; Mcdonald et al., 2006; McKenna et al., 2012). Large commercial ships have been found to generate noise through all frequency bands, with a majority of energy concentrated below 1000Hz. Each ship generates a unique acoustic signature dependent on its characteristics (e.g. size and load) and
operational conditions (e.g. Ross, 1976; McKenna, et al., 2012). Higher shipping density in the sea can result in larger noise levels (Ross, 2005), particularly near the major ports and heavily used shipping lanes (Wright, 2008). There has also been a growing concern that such low-frequency and continuous anthropogenic sound can cause chronic effects on marine species (e.g. Southall et al., 2007; Ellison et al., 2011), which are difficult to observe and measure in comparison to acute events. Understanding the characteristics of shipping noise and its temporal and spatial distribution is, thus, of particular importance to the protection of marine life.

The AIS is a collision avoidance and vessel-tracking system for large commercial ships which operates on the VHF radio bandwidth. It is compulsory for vessels exceeding 299 gross tonnes to install the AIS transceivers in accordance with the international convention for the Safety Of Life At Sea (SOLAS, 2005). A centralised database contains information regarding the properties of vessels and operational conditions, allowing exchange of data for identifying and locating nearby vessels. The data are now widely utilised for shipping noise studies (e.g. Hatch et al., 2008; Merchant et al., 2012) as they provide positions of vessel movements in the sea, vessel properties and operational speeds which can be used to estimate the source level of noise generated by a ship.

Traditional marine mammal management focused primarily on highly-intense and short-term impacts from lethal events (e.g. vessel collisions and strandings), but the chronic effects from long-term disturbance or sub-lethal effects from less intensity noise received less attention (Wright et al., 2011). An elevated shipping noise level raises the concern of chronic effects on marine animals, especially in biologically important coastal waters. These negative effects can be categorised specifically as masking of biologically significant sound, behavioural disturbance and physiological stress (Southall et al., 2007; Wright, 2008). Noise can overlap
the biologically important sounds over the functional frequency bandwidth of animals, therefore affecting the ability to detect cues from conspecifics, echolocation from prey, echoes aided for navigation. Many marine mammals (e.g. baleen whales and seals) and fish have been known to be extremely sensitive to low frequency shipping noise as they emit similar low frequency sound for breeding, foraging and navigation; and other cetaceans (e.g. dolphins and porpoises) may also encounter masking when staying close to the proximity of shipping lanes (Southall, et al., 2007; Wright, 2008).

There is growing evidence that noise from ships has the potential to mask the communication (e.g. Erbe, 2002; Clark et al., 2009), resulting in changes in behaviour such as the fluke rate and dive depth (e.g. Nowacek et al., 2004), increasing the amplitude of calls (Parks et al., 2011), reducing calling rates (e.g. Lesage et al., 1999) and note duration (e.g. Castellote et al., 2012), and leading to physiological stress (e.g. Rolland et al., 2012; Wale et al., 2013). These responses have great influence on energetic cost to animals for survival, hence affecting navigation, foraging and reproductive activities in marine habitats.

MCZs are designed to protect the diversity of nationally rare, threatened and representative habitats and species. The Celtic Sea is situated within the South-West Deeps MCZ, one of twenty seven MCZs in UK which have been designated on 21 November 2013 (MCZs report, 2013). This site is located to the southwest of England, protecting a total area of about 1800km². This site which is comprised of different sediments forms various habitats that support a large number of species, such as small burrowing worms living within the sediments and crustaceans that live on the sediment surface (McBreen et al., 2011). The Celtic Sea sandbanks are also protected by this MCZ due to their geological importance. Many activities (e.g. renewable energy) within the MCZs have been regulated through marine
licences and an environmental impact assessment is required before a licence decision is made.

The MSFD outlines the legislative framework to the management of anthropogenic activities in order to achieve the Good Environmental Status by 2020 in the European marine environment. With regards to underwater noise, marine animals that are exposed to human-made sound can be affected adversely in the forms of acute effect (short term) and chronic effect (long term). A report (Van Der Graaf et al., 2012) by MSFD has proposed the guidance that is used to help implement two important indicators: low and mid frequency impulsive noise and ambient noise. The preliminary monitoring scheme for the ambient noise suggested by Van Der Graaf et al. (2012) is to identify trends in the sound level within 1/3 octave bands 63 and 125Hz centre frequencies, typical values where peak of shipping noise occurs. The report also indicates clearly that noise modelling is an essential complement to measurements, especially to examine the variability of the sound distribution resulted from large scale changes in climates, oceans and other factors. To meet the requirements by MSFD, the UK Marine Strategy (UKMS, 2014) has been developed to establish and implement a monitoring programme to measure progress towards achieving Good Environment Status, provided with detailed monitoring programmes for 11 categories. Importantly, there is insufficient monitoring data to support an assessment of current ambient noise levels or their impact on marine animal populations in UK as stated by UKMS (2014).
Chapter 3

3 Methodology

This chapter presents the main methods utilised by this study, including the ocean modelling, the coupled ocean-acoustic modelling and some basic statistical methods used for data processing, while other approaches related to the specific purpose and subject are illustrated in the corresponding chapters.

3.1 The ocean model: POLCOMS

3.1.1 Model description

The ocean model used for this study is the Proudman Oceanographic Laboratory Coastal Ocean Modelling System (POLCOMS) described by Holt and James, (2001). It is a baroclinic three-dimensional primitive equation finite difference model solving motion equations with the incompressible, hydrostatic and Boussinesq approximations. The model uses the Arakawa (1972) B-grid solving for eastward and northward velocities, elevation, potential temperature and salinity and turbulent kinetic energy. It uses the Piecewise Parabolic Method (Collella and Woodward, 1984; James, 1996) for advection, the Laplacian diffusion with Smagorinsky (1963) algorithm for horizontal diffusion and the k-ε turbulence closure scheme. As described previously (see section 2.2), this model has been developed intensively for many years and there is a track record of its development. The details of
model physics can be found in Holt et al. (2001), the POLCOMS technical description\textsuperscript{1} and the POLCOMS user guide\textsuperscript{2}, which are also summarised in Appendix 1.

3.1.2 Model set-up

This section presents the detailed forcing, methods and model code changes used to set up the model in the Celtic Sea. The flow chart (see Fig. 3.1) acting as the backbone throughout the section of model set-up summarises the general procedures for manipulating POLCOMS. The vertical column (rectangular boxes without shading) on the right hand side of Fig. 3.1 describes the primary axis of procedures while others illustrate the auxiliary data processes for each procedure. The specific description for each procedure shown in Fig 3.1 will be described in detail. In the figure, initial data and the subsequently manipulated data are shown as the diamond-shaped boxes whilst processes are expressed in rectangular boxes. It is worth noting that this flow chart gives specific procedures suitable for this study rather than universal applications of POLCOMS, thus the methods used in this study would require modifications for their use on different projects. In order to differentiate files of POLCOMS, a typographic convention is defined as follows: the names of model code files, subroutines and modules are highlighted in bold font.

The model set-up of POLCOMS for this study is based on the standard model developed by Shapiro (2011), with following differences: (i) the model domain has been extended westwards from 7.00°W to 7.90°W to cover a larger geographic area in the Celtic Sea. The northwest corner reaches the southeast coast of Ireland at a water depth of 5m. The input data including bathymetry, surface forcing, initial fields for temperature and salinity and open boundary are recalculated in order to be compatible with the new model domain. The sources

\textsuperscript{1}available online at http://cobs.pol.ac.uk/modl/metfcst/POLCOMS_DOCUMENTATION/node2.html
\textsuperscript{2}available online at http://cobs.noc.ac.uk/modl/polcoms/POLCOMS_user_guide.pdf
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Fig. 3.1 Flow chart summarising the procedures for manipulating POLCOMS
of the bathymetry data, initial fields, open boundary temperature and salinity are the same as that in Shapiro (2011), but the surface forcing used by this study is extracted from different databases in order to perform the sensitivity studies while only NCEP data is used by Shapiro (2011). (ii) As tides are the dominant forcing in forming the seasonal features in the Celtic Sea (Pingree, 1980), it is important to simulate accurately the tidal currents using an ocean model. The tidal forcing at the open boundaries has been changed to be compatible with the 1/30° x 1/30° TPXO 7.2 Regional Inverse Tide Model developed by Oregon State University (Egbert et al., 2010). Two shallow water constituents (MS₄ and MN₄: shallow water quarter diurnal constituents) that are not adopted by Shapiro (2011) have been added to the open tidal boundary forcing. The model code for the calculation of the barotropic tides is modified and the predicted tidal velocities are validated. (iii) Full model validations which have not been performed by Shapiro (2011) are carried out in this study by comparing to intensive observational data. (iv) With regards to the horizontal diffusion, the free constant for the horizontal turbulent diffusivity calculated using Smagorinsky (1963) scheme is chosen to be 0 rather than 0.2 used by Shapiro (2011) in order to improve the representation of the subsurface temperature fronts.

The procedure of replacing the previous tidal open boundary conditions of POLCOMS with the TPXO 7.2 Regional Inverse Tide Model will be described in detail in the next section while technical descriptions for other procedures (see Fig. 3.1) are introduced in Appendix 2 as a supplementary to this section. Instead, a brief summary of the model set-up is described in this section as follows:

This study focuses on an area in the Celtic Sea between 50.08°N to 51.83°N and 7.90°W to 4.00°W, surrounded by the north Cornish coast, the Bristol Channel, the south St. George’s Channel and the west and south opening boundaries. Fig. 1.1 shows the model domain as
covered by the red rectangular box. In order to exclude wetting and drying effects, numerical calculations are performed in the areas only deeper than 5m. The model uses an $s$-coordinate system with 30 vertical layers with parameters $h_c = 200$, $\rho = 8$ and $b = 0.05$ (see Appendix 2 for details), and a horizontal resolution of ~2km formulated with the Arakawa (1972) B-grid. The bathymetry was acquired from the ETOPO-2 database (ETOPO2, 2006) which has a resolution of 2 minutes in both latitude and longitude directions corresponding to latitudinal ~4km and longitudinal ~2km respectively, and smoothed via the application of a 3 x 3 low pass matrix filter. The full model simulation was run using meteorological forcing mainly from NCEP-II data set (NCEP-DOE Reanalysis-II, 2013), including following 7 variables: $u$ and $v$ components of the wind at 10m above sea surface, air temperature at 2m, atmospheric pressure, relative humidity, cloudiness, and precipitation rate. Downward solar radiation was calculated by the model from astronomical data and cloudiness. The initial fields of temperature and salinity and boundary conditions were extracted from the World Ocean Database (Boyer et al., 2004). Simulations were initialised using the ‘semi-diagnostic adjustment method’ (Enriquez et al., 2005; Sarkisian and Sündermann, 2009), i.e. by running the model with no meteorological forcing and river discharges with frozen initial temperature and salinity distributions (Enriquez et al., 2005). This allows the currents to achieve equilibrium with the density field.

Heat fluxes are calculated in POLCOMS from meteorological data using a version of bulk formulae by Elliott and Clarke (1991), following Gill (1982), requiring the following four variables: air temperature, atmospheric pressure, relative humidity and total cloud cover. The surface forcing data, whatever the resolution is, are interpolated onto the model grid. The inward and outgoing heat fluxes are treated separately in POLCOMS. The cooling process at the sea surface from the outgoing flux is parametrised by equation (3.1):
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\begin{equation}
T^{p+1} = T^p + \frac{q_I}{c_p \rho} \times \frac{\Delta t}{\Delta z}
\end{equation}

(3.1)

where \(T^p\) is the surface temperature at model time \(p\) and \(T^{p+1}\) is the new one at the time step. \(q_I\) is the heat flux leaving surface and \(c_p\) is the specific heat of sea water. \(\Delta t\) is the time step and \(\Delta z\) is the depth of the surface model grid. \(\rho\) is a reference density of seawater. The new surface temperature is updated by such iteration algorithm.

The inward heat flux is transferred to the seawater through the mechanism:

\begin{equation}
T_k^{p+1} = T_k^p + \frac{q_{in}}{c_p \rho} \times \frac{\Delta t}{\Delta z} \times (e^{\varepsilon z_k} - e^{\varepsilon z_{k-1}})
\end{equation}

(3.2)

where \(q_{in}\) is the incoming heat flux calculated using the bulk formulae. \(T_k^{p+1}\) and \(T_k^p\) are the new and old temperature at a water depth of \(k\). \(\Delta t\) is the time step and \(\Delta z\) is the distance between two levels of vertical coordinate system. \(z_k\) and \(z_{k-1}\) are depths of level \(k\) and \(k-1\) respectively. \(\varepsilon\) is the transmissivity.

The wind stress in POLCOMS are calculated based on 10m wind speeds using:

\begin{equation}
\tau_i = \frac{\rho_a}{\rho} (0.63 + 0.066) \times 10^{-3} W u_i
\end{equation}

(3.3)

where \(u_i\) is the wind speed components and \(\tau_i\) is their corresponding stresses. \(W\) represents the scalar of the wind speed and \(\rho_a\) is the air density.

The precipitation rate is also required to adjust the salinity field.

The south boundary of the model is 50.08°N (see Fig. 1.1) which is not extended to cover the region of the shelf break. The shelf break is always of particular interest to sound propagation due to its rapid changes in bathymetry. This is limited by the computer power as this POLCOMS model used in this study is configured under Windows operational system that
does not support parallel computation. However, a new version of NEMO model (Wobus et al., 2013) has been developed recently to be compatible with the parallel environment running on high performance computers of Plymouth University. The POLCOMS model uses the traditional $s$ coordinates described by Song and Haidvogel (1994) while the hybrid enveloped vertical discretisation of NEMO developed by Shapiro et al. (2013) shows remarkable improvements in the areas of steep topography. Another advantage of the NEMO model is the non-linear free surface with vertical piecewise parabolic method (vPPM) for vertical advection that is not used by POLCOMS. In term of the parameterisation of the surface forcing, the POLCOMS model used by this study utilises the bulk formulas from Elliott and Clarke (1991), following Gill (1982) whereas NEMO contains a number of bulk formulas including the COARE v3 algorithm (Fairall et al., 2003). A study from O’Neill et al. (2012) has shown that a coarse resolution 7km NEMO model is able to predict the temperature and salinity fields in Liverpool Bay in the equivalent confidence to a fine resolution 1.8km POLCOMS model. Although NEMO is being continuously improved and used by a large community, the POLCOMS model is still capable for this study due to its good features for shelf sea modelling.

SST plays an important role in the atmosphere-ocean system as it is of great influence on the regulation of the air-sea interactions (Kantha, 2000). Air masses in the atmosphere system are affected significantly by SST whilst atmosphere variables also have strong impacts on SST in various scales. Each component of the atmosphere variables has different capability in driving the SST seasonal cycle in global ocean. A study from (Kara et al., 2009) has shown that shortwave radiation is the most influential variable controlling the seasonal cycle of SST over 34.3% of the global ocean and wind speed is the second most important variable (27.2%). Ocean response is extremely sensitive to SST variations, which can further affect the density structures and the currents patterns. In models, the accuracy of the modelled SST
is a good indicator that reflects the performance of the model. The performance of POLCOMS has been intensively assessed by Holt et al. (2005) where a 7km model has been implemented in the northwest European continental shelf. The model is forced by 6 hourly ECMWF meteorological data using the bulk formulae described by Holt and James (1999a). The modelled overall SST RMS error in their study is 1.00°C and 0.09°C for the mean, which is similar to that (~1.20°C and ~0.13°C) in Holt and James (2001) where the 3 hourly U.K. Met Office weather prediction model date are forced to the model using the bulk formulae as the same as used in this study. Both a 1.8km POLCOMS model and a 7km NEMO model which are forced by the U.K. Met Office weather prediction model date, but using the COARE v3 algorithm (Fairall et al., 2003), are applied in Liverpool Bay to predict the temperature and salinity fields, giving an equivalent RMS errors in SST by ~1.38°C and 1.40°C respectively.

3.1.3 Defining open boundary tide forcing

Barotropic tidal elevation and currents of POLCOMS are forced using the Flather open boundary condition, allowing adjustment to internally generated waves. In the previous version of POLCOMS, the total tides are the sum of 15 harmonic analysis constituents (see table 3.1), which are usually extracted from a larger scale POLCOMS model. This model was being running at the Proudman Oceanographic Laboratory (POL) and the tidal constituent data are usually provided by POL under permission. The data representing these 15 constituents are the amplitudes and phase speeds, which are used to apply nodal factors and date corrections to give the correct tidal phase for the specified date. The calculation of nodal factors and date corrections is conducted internally in POLCOMS while the data of the amplitude and phase speed need to be prepared forehand as an external input file.
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<table>
<thead>
<tr>
<th>Table 3.1 Tidal constituents used by POLCOMS and TPXO 7.2 Regional Inverse Tide Model</th>
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<tbody>
<tr>
<td>Models</td>
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<tr>
<td>POLCOMS</td>
</tr>
<tr>
<td>TPXO 7.2 Regional Inverse Tide Model</td>
</tr>
</tbody>
</table>

As simulations in this study are carried out in a shallow water of the European continental shelf, the non-linearities of tidal constituents, such as the overtide M₄ or the compound tide MS₄ become important. An accurate shelf tidal model is, hence, indispensable to this study since tides are the dominant forcing of the seasonal stratification and associated tidal fronts in the Celtic Sea (Pingree, 1980). The European shelf tide model (hereafter referred to ES2008) from Oregon State University, with a resolution of 1/30° in both directions, is selected to provide tidal constituents for POLCOMS due to its easier accessibility and improved fits to measurements. ES2008 (Egbert et al., 2010) is designed to predict tides, as well as extract the constituents for the European continental shelf waters. Using a new variational data assimilation scheme for compound tides and overides, prior solutions for M₄, MS₄ and MN₄ computed using inverse solutions for M₂, S₂, and N₂ dramatically improve fits to validation tide gauges relative to unconstrained forward solutions (Egbert et al., 2010). This model considers primary and secondary tidal constituents as a coupled problem using a simple linearized perturbation theory for weak interactions of the dominant primary constituents, which gives the RMS error less than ~2cm for the validation of elevation.

ES2008 contains 11 tidal constituents while it is 15 for the previous POLCOMS (see table 3.1). In order to make POLCOMS compatible with ES2008 and maintain the original features of POLCOMS, the model was modified with the capability that POLCOMS can use both tidal boundary conditions. The procedures of coupling the ES2008 tidal model with POLCOMS are described in the following three sections.
1) Physical background of tidal calculation

The currents and elevations of each constituent (U, V, and Z respectively) are expressed by:

\[
\{X_n\} = H_n f_n \cos[\tilde{\sigma}t - g_n + (V_n + u_n)]
\]  

(3.4)

where \(H_n\): The average amplitude. \(f_n\): The nodal factor. \(\tilde{\sigma}\): The angular speed (degree/hour). \(g_n\): The local phase lag. \(V_n\): The tidal phase of 0 GMT on the start day of running. \(u_n\): The nodal angle. \(X_n\) represents the tidal velocities or elevations as the case may be.

Expand equation 3.4 as following:

\[
\{X_n\} = H_n f_n [\cos(\tilde{\sigma}t - g_n) \cos(V_n + u_n) - \sin(\tilde{\sigma}t - g_n) \sin(V_n + u_n)]
\]

\[
= H_n f_n [\cos(\tilde{\sigma}t) \cos(g_n) \cos(V_n + u_n) + \sin(\tilde{\sigma}t) \sin(g_n) \cos(V_n + u_n) - \sin(\tilde{\sigma}t) \cos(g_n) \sin(V_n + u_n) + \cos(\tilde{\sigma}t) \sin(g_n) \sin(V_n + u_n)]
\]

\[
= H_n \cos(g_n) \times f_n \cos(V_n + u_n) \times \cos(\tilde{\sigma}t) + H_n \sin(g_n) \times f_n \sin(V_n + u_n) \times \cos(\tilde{\sigma}t) +
\]

\[
H_n \sin(g_n) \times f_n \cos(V_n + u_n) \times \sin(\tilde{\sigma}t) - H_n \cos(g_n) \times f_n \sin(V_n + u_n) \times \sin(\tilde{\sigma}t)
\]  

(3.5)

Underlined terms in equation 3.2 represent the parameters derived from the constituent data, while the others are calculated internally in the subroutine \texttt{tideset} of POLCOMS. Afterwards, POLCOMS couples the underlined terms prepared as external input data and the other terms (nodal factors and phase corrections) to calculate \(X_n\). The underlined terms contain the cosine, sine components and \(\tilde{\sigma}\) (angular speed) of each constituent, which correspond to the variables: (\(z1\), \(u1\) and \(v1\), \(z2\), \(u2\) and \(v2\)) and \texttt{sigma} in the subroutine \texttt{tideset}. Fig. 3.2 shows the flow chart that summarises the procedures of tidal calculations at open boundary
points in POLCOMS. The dashed rectangular box in Fig. 3.2 covers the calculations occurred in the model whilst the others are processes for preparing the external input file.

2) Modifications of model internal calculations

There are two shallow water constituents (MS$_4$ and MN$_4$: shallow water quarter diurnal constituents), which are not adopted by the previous POLCOMS (see Table 3.1). New subroutines which conduct calculations of nodal factors and phase corrections (terms with no underlines in equation 3.5) for MN$_4$ and MS$_4$ are necessary to make ES2008 compatible with POLCOMS. Following Fig. 3.2, the start time is given by an external file and the calculations of astronomical arguments are kept intact. The computational equations and rules of nodal factors $f_n$ and $u_n$ refer to the harmonic constituents analysis edition 1.0 (2009) from the UK Hydrographic Office (see Appendix 3) whilst the method of computing $v_n$: tidal phase of 0 GMT on the start day of running is described in Kantha (2000). The resulting new subroutines are named as $ufset_{ES2008}$ and $vset_{ES2008}$ respectively. To optimise the entire model, the new version is designated to be compatible with both the previous POLCOMS tide model and the ES2008 model. A control file (header Tide.h) is added to choose which tide model will be used. If the directive `ifdef OSU_ES2008_tide` is active, the tide calculations will be carried out following different subroutines and definitions which correspond to the ES2008 model.

3) Prepare the external input file

Apart from changes made to the model, the input data are also required to be consistent to the new subroutines. The constituent data are extracted from the ES2008 model to calculate the underlined terms as marked in equation 3.5. Procedures of producing the input data are introduced as follows:
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a) Download ES2008 model (1/30°) from Oregon State University website http://volkov.oce.orst.edu/tides/ES.html

b) Extract the compressed file in which it contains four files listed as below:

- Tidal model control file (ASCII, supplied with the model): Model_ES2008
- Elevation model file name: hf.ES2008.out
- Transport model file name: uv.ES2008.out
- Bathymetry grid file name: gridES

c) Download the Tide Model Driver (TMD)

The TMD toolbox includes a user-friendly graphical user interface (GUI). It is a MATLAB® package for accessing the harmonic constituents for the Oregon State University family of high-latitude tide models, and for making predictions of tide elevation and currents. TMD includes two components: (i) a GUI for quickly browsing tide fields, zooming in on regions of interest, and selecting points and time ranges for predictions of specific variables; and (ii) a set of scripts for accessing tide fields and making predictions. The GUI of TMD makes this driver easy to access and a detailed description about how to use it was included in the compressed package. Users can follow the instructions interpreted in the read-me file to either make the tide predictions or extract harmonic constituents.

d) Create the control file

A control file (tidin_run_ES2008.dat) should contain the start time of a simulation, the number of tidal constituents, and flags (1 or 0) designed to choose which constituent is to be used.

e) Extract tidal constituents for each boundary point
Firstly, generate the open boundary point files, which include boundary point information, for instance the number of points and the location of each point. Note that the number, latitude and longitude are different between UV and Z because of the usage of B-Grid. Subsequently combine these files and use MATLAB® to produce the input data, giving the boundary locations for the TMD driver. Finally run the main script (tmd.m) of TMD to extract constituents from the ES2008 model.

Calculate the cosine and sine components marked with underlines in equation 3.2 for each constituent and rearrange the order of the data to be compatible with the internal calculations. This is achieved by using FORTRAN®. The procedures of producing the boundary tidal data are summarised in the flow chart (Fig. 3.3).
Read in $H_n \cos(g_n)$ and $H_n \sin(g_n)$

Input files containing $H_n \cos(g_n)$; $H_n \sin(g_n)$; $\sigma$

Read in angular speed $\sigma$
Calculate $\cos(\sigma t)$

Start time

$\frac{f_n \sin(V_n + u_n)}{f_n \cos(V_n + u_n)}$

Calculate phase of 0 GMT; $V_n$

Calculate nodal parameters; $f_n$ and $u_n$

Calculate astronomical arguments for tides

$s$, $h$, $p$, $en$, $p1$

$s$: Mean longitude of Moon
$h$: Mean longitude of Sun
$p$: Mean longitude of lunar perigee
$en$: Mean longitude of ascending node
$p1$: Mean longitude of solar perigee (perihelion)

**Fig. 3.2** Flow chart summarising the procedures of tidal calculation at open boundary points. Dashed rectangular box covers the calculation performed internally in POLCOMS.
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Fig. 3.3 Flow chart summarising the procedures of producing open boundary tidal data
3.2 The acoustic model: HARCAM

The acoustic model used for this study is HARCAM, which is a software engine for the Naval Tactical Decision Aid WADER-32. It has been validated by the UK Royal Navy in 2009 and has been utilised operationally by the Royal Navy in both deep and shallow seas. WADER32 system is the mandated acoustic model for use throughout the UK Royal Navy and is used by many other navies and research establishments worldwide (see http://bdec-online.com/bd-cat36/c360003.pdf).

Underwater sound propagation is the basis combined with several loss mechanisms. In accordance with the underlying physics, the propagation model is generally classified as a ray theory model, a normal mode model, or using wavenumber integration techniques and parabolic equations. The ray theory model is the most straightforward method to solve the wave equation, but is subject to the approximation in its underlying physics (see the details in the next section). This approach is applicable when scales of variability of oceanic processes are larger than the sound wave length. The ray model is, hence, generally used for higher frequency. It is, however, computational fast and can give the physical structure of sound propagation paths. The parabolic equation model is reliable model that is able to handle range-dependent problem, but it is not practical computationally when the frequency is higher than 500Hz (Etter, 2001). The normal mode model and wavenumber integration techniques are both applicable physically and practical computationally, they are, however, not suitable for range-dependent problem. The HARCAM model is a combination of ray model and parabolic model, which are designed for high frequency and low frequency range-dependent problems respectively. The optimised separation for the frequency is 150Hz which is verified by the Royal Navy (see http://bdec-online.com/bd-cat36/c360003.pdf). A unique feature of HARCAM is that the ray model is used to provide the surface and absorption data to correct
the parabolic model outputs between 250-1000Hz where the parabolic model is unable to provide the necessary data. In the next section only the ray theory and parabolic equation are introduced as HARCAM is based on these two theories.

3.2.1 Basic concepts

Sound Speed

The transmission of sound waves in waters is governed by variations of the sound speed, which is relative to the physical properties of water columns to determine the path of propagation, and the chemical composition of the sea to determine the energy loss (Etter, 2001). The sound speed is controlled by spatial and temporal variations of temperature, salinity and pressure (Urick, 1982). In a stratified medium, such as the ocean, acoustic beams propagate not along the line of sight, but a curved trajectory due to multiple refractions and experience multiple reflections from the seabed and sea surface (Brekhovskikh et al., 2003). A standard equation of describing the relationship between sound speed and three water properties was introduced by Mackenzie (1981) as shown by equation 3.6:

\[
c = 1448.96 + 4.519T - 5.304 \times 10^{-2}T^2 + 2.374 \times 10^{-4}T^3 + 1.34(S - 35) + 1.63 \times 10^{-2}D + 1.675 \times 10^{-7}D^2 - 1.025 \times 10^{-2}T(S - 35) - 7.139 \times 10^{-13}TD^3 \quad (3.6)
\]

where \(T\) is the temperature (°C), \(S\) is the salinity (PSU) and \(D\) is the water depth in metres. A quantitative relation between sound speed and three water properties was described by Etter (2003) shown as follows:

- Increase in sound speed by ~3.6ms\(^{-1}\) per 1°C increase in temperature
- Increase in sound speed by ~1.4ms\(^{-1}\) per 1PSU increase in salinity
- Increase in sound speed by ~1.7ms\(^{-1}\) per 100m increase in depth
In deep waters temperature variations in the vertical direction are usually of greater influence to the sound speed in the top 1000m, below which depth becomes dominant, while vertical variations of salinity are often negligible except in the areas of excessive evaporation and precipitation (Etter, 2003). However, in shallow waters the salinity may account for the primary impacts on the sound energy fluctuation, the areas with strong river run-off for example.

**Transmission loss (TL)**

TL is an important concept in underwater acoustics. It is a measure of accumulated decrease in acoustic intensity as underwater sound propagates outwards from a source. It is defined by equation 3.4 and the unit is in decibel (dB):

\[
TL = 10 \log_{10} \frac{I}{I_r}
\]  

(3.7)

where \(I\) is the sound intensity to be measured at a point, and the reference intensity \(I_r\) is measured at a point 1m away from the source. Alternatively, TL can be expressed by the sound pressure (see equation 3.8), as the sound intensity in a plane wave is proportional to the square of the pressure amplitude. It is used widely in acoustics since sound pressure can be measured directly.

\[
TL = 20 \log_{10} \frac{P}{P_r}
\]  

(3.8)

Acoustic energy is lost generally through two major ways, namely spreading and absorption. Two geometric spreading models are widely used to simply estimate the TL and defined as:

The spherical spreading model:

\[
TL = 20 \log_{10} r
\]  

(3.9)
where \( r \) is the propagation distance, and the cylindrical spreading model:

\[
TL = 10 \log_{10} r 
\]  

(3.10)

Both models are a time-depth-independent logarithm dispersion. The former assumes that energy radiated from a point source is equally distributed over the surface area of a sphere surrounding. For such cases the energy does not encounter any boundary interactions (in deep oceans for example). The latter represents the transmission loss in a waveguide that has plane upper and lower boundaries (shallow water for instance). Two models are utilised broadly for different applications. For example, biological regulatory bodies use these two models to assess the likely impacts of anthropogenic sound by estimating the sound propagation (Southall et al., 2007). It can be seen from equation 3.9 and 3.10 that these two models are independent of environmental variations, which are, however, highly important to the sound propagation in shallow waters. Comparisons in the level of TL predicted by HARCAM and spreading models are, thus, performed in order to evaluate the difference in predicting the sound exposure in shallow waters, with the results being presented in chapter 6.

### 3.2.2 Propagation models

The fundamental theory underlying all mathematical models of acoustic propagation is the wave equation, which is derived from the equations of state (adiabatic), continuity and motion (Etter, 2003). For most applications, the wave equation (3.11) is simplified, hyperbolic, second-order and time-dependent.

\[
\nabla^2 \phi = \frac{1}{c^2} \frac{\partial^2 \phi}{\partial t^2} 
\]  

(3.11)

where \( \nabla^2 \) is the Laplacian operator. \( \phi \) is the potential function. \( c \) is the speed of sound and \( t \) is time. The wave equation is normally solved in the frequency domain. By applying a Fourier
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Transform to equation 3.11 and using the variable separation method, we can obtain the Helmholtz equation for a homogeneous medium:

\[ \nabla^2 \phi + K^2 \phi = 0 \]  \hspace{1cm} (3.12)

where \( K = \frac{\omega}{c} \), \( \omega \) is the angular frequency of source and \( c \) is the sound speed. By doing so the original wave equation is reduced to be time-independent. Different solutions to equation 3.12 yield different propagation models.

In shallow waters, for the appropriate solution of the Helmholtz equation, boundary conditions need to be specified. At the sea surface, the pressure is defined to zero: \( p = 0 \). At the interface between the sea water and sediment layer, the continuity of pressure and the normal component of the particle velocity are given by:

\[ p_1 = p_2 + \frac{1}{\rho_1} \frac{\partial p_1}{\partial t} = \frac{1}{\rho_2} \frac{\partial p_2}{\partial t} \]  \hspace{1cm} (3.13)

where \( p_1 \) and \( p_2 \) are sound pressures of water and sediment at the bottom interface respectively.

**Ray theory propagation model**

Ray theory propagation is a geometric approximation to solve equation 3.9, hence TL. Similarly, the Helmholtz equation for an inhomogeneous medium in cylindrical coordinates can be expressed by:

\[ \nabla^2 \phi(r, z) + K^2(r, z)\phi(r, z) = -\delta(r - r_s)\delta(z - z_s) \]  \hspace{1cm} (3.14)

where \( r \) is the propagation range and \( z \) is the water depth. \( \delta \) is the Dirac delta function. The subscript ‘\( s \)’ denotes the coordinates of source. Coefficient \( K^2(r, z) = \frac{\omega}{c(r, z)} \) is the
representation of range-dependent environment. The solution for φ is assumed to be the form:

\[ \phi = F(x, y, z)e^{-iG(x, y, z)} \]  

(3.15)

where F represents a pressure amplitude function and G a phase function. The exponential term allows for rapid variations as a function of range while F is a more slowly varying envelope incorporating the geometrical spreading and loss mechanisms. Substituting equation 3.15 to equation 3.14 and separating the real and imaginary terms give:

\[ \frac{1}{F} \nabla^2 F - (\nabla G)^2 + K^2 = 0 \]  

(3.16)

\[ 2(\nabla F \cdot \nabla G) + F \nabla^2 G = 0 \]  

(3.17)

Equation 3.16 defines the geometry of the rays whilst equation 3.17 determines the wave amplitudes. This separation is based on the approximation that the fractional change of sound speed gradient over a wavelength is small compared with the gradient \( c/\lambda \) (Etter, 2003) where \( \lambda \) is the wave length. Mathematically: \( \frac{1}{F} \nabla^2 F \ll K^2 \), which means that the sound speed must not change over one wavelength. Under this approximation equation 3.16 can be reduced to the eikonal equation:

\[ (\nabla G)^2 = K^2 \]  

(3.18)

Surfaces with constant phase \( G \) are the wavefronts and the eigenrays to wavefronts are the rays. The trajectories are perpendicular to the surfaces with constant phase. The amplitude of the sound field at any point can be calculated from the density of rays using equation 3.17.

This method is computationally efficient and able to deal with range-dependent problems. The ray traces can give the physical structure of sound propagation paths, and thus are helpful
in describing how the sound field redistributes itself over long propagation distances. However, it does not include diffraction effects which are significant at low frequencies.

**Parabolic equation model**

The parabolic equation (PE) method to solve the wave equation is derived by assuming that energy propagates at speeds close to a reference speed (Etter, 2003). It factors an operator to obtain an outgoing wave equation that can be solved as an initial-value problem in a range. By dividing a range-dependent medium into a sequence of small steps, each step can be treated as the initial-value problem. The entire region is then obtained by ‘marching’ each small step based on energy conservation.

The propagation equation for a point source is also given by the Helmholtz equation:

\[
\nabla^2 \phi + k_0^2 n^2 \phi = 0
\]

(3.19)

where \(k_0\) is the reference wave number, \(k_0 = \omega/c_0\) and \(c_0\) is the reference sound speed. \(n\) is the refraction index \(c/c_0\). For each step \(n\) is taken to be a constant. By neglecting azimuthal angle dependence (reduced to a 2D problem), equation 3.19 can be expressed in cylindrical coordinates as:

\[
\frac{\partial^2 \phi}{\partial r^2} + \frac{1}{r} \frac{\partial \phi}{\partial r} + \frac{\partial^2 \phi}{\partial z^2} + k_0^2 n^2 \phi = 0
\]

(3.20)

where \(r\) is the horizontal propagation range and \(z\) is the water depth. Assume a solution in the form of:

\[
\phi = \Psi(r, z) \cdot S(r)
\]

(3.21)

substitute equation 3.21 to equation 3.20 and use \(k_0^2\) as a separation constant. Then we obtain two equations:
and

\[
\left[ \frac{\partial^2 S}{\partial r^2} + \frac{1}{r} \frac{\partial S}{\partial r} \right] = -Sk_0^2 \tag{3.22}
\]

Equation 3.22 is a Bessel equation for outgoing waves and we take the outgoing solution given by the zero-order Hankel function:

\[ S = H_0^{(1)}(k_0r) \tag{3.24} \]

With the far-field approximation we have the form of asymptotic expansion for large arguments:

\[ S \approx \frac{2}{\sqrt{\pi k_0 r}} e^{i(k_0 r - \pi/4)} \tag{3.25} \]

With the paraxial approximation \( \frac{\partial^2 \psi}{\partial r^2} \ll 2k_0 \frac{\partial \psi}{\partial r} \) and by substituting equation 3.25 to equation 3.23, equation 3.23 can be further simplified to:

\[ \frac{\partial^2 \psi}{\partial z^2} + 2i k_0 \frac{\partial \psi}{\partial r} + k_0^2 (n^2 - 1) \psi = 0 \tag{3.26} \]

Equation 3.26 is the parabolic wave equation which has been reduced to a first-order differential equation in the horizontal range \( r \). It can be numerically solved by using ‘marching solution’ by giving the initial field (Jensen et al., 2011). A common method is to initialise the start field by using a normal mode representation.

The PE model, treated as an initial-value problem in range and assuming that propagation is along a 2D horizontal plane, produces a very accurate TL solution with range. It is, however, extremely computationally intensive for practical applications, especially at frequencies over
500Hz above which it is not applicable for most active sonars (Etter, 2003). Another weakness of PE is that those models cannot solve interactions at boundaries very well, instead using a simplified reflection parameterisation to describe the boundary processes (Hodgson, 2011).

### 3.2.3 Geoacoustic parameters

It is well known that in shallow water the bottom reflection loss and absorption are dominant loss mechanisms and acoustic propagation modelling requires accurate representation of geoacoustic properties of the seabed (Hamilton et al., 1982). The ocean bottom sediments are often modelled as fluid layers which only support a compressional wave (Jensen et al., 2011). Reflectivity is the measure of reflection when sound propagates through the interface of two mediums. The critical grazing angle is the angle separation between perfect reflection and reflection with energy loss, below which sound does not encounter energy loss. Above the critical grazing angle bottom loss occurs which is a non-linear function of grazing angle.

Real ocean bottoms are complex layered structures. A complete geoacoustic model must contain the sediments and basement properties through the effective acoustic penetration depth, depending on the frequency of the source (Jensen et al., 2011). At high frequencies details of the bottom are required through the upper few metres as sound energy can be absorbed rapidly. However, at low frequencies information must be provided over the whole bottom and even the underlying rocks.

The seabed composition mapping of UK waters was conducted during the UKSeaMap 2010 project which combines the sediment data collected in a series of surveys from 1967 to 2009 (McBreen, et al., 2011), providing detailed maps of sediment distribution in the Celtic Sea. The corresponding geoacoustic parameters for HARCAM were taken from the study by Hamilton (1980) for muddy sand, sand and gravelly sand types of seabed, see Table 3.2.
CHAPTER 3. METHODOLOGY

These parameters were widely used in underwater acoustics, e.g. Lermusiaux et al. (2010); Holland and Dettmer (2013). Geoacoustic data from Hamilton (1980) do not, however, cover the sandy gravel, and the parameters for this type of seabed were taken from NATO Research Centre sonar acoustic handbook (NURC, 2008). The total sediment thickness distribution in the Celtic Sea was extracted from the NOAA National Geophysical Data Center (Divins, 2003). All geoacoustic parameters for HARCAM are summarised in Table 3.2. This ensures that HARCAM can capture the range-dependent parameters of seabed in calculating the transmission loss, rather than a single bottom type which were adopted in previous studies (e.g. Xu et al., 2009; Lam et al., 2009). Those geoacoustic parameters in Table 3.2 are conditioned to HARCAM in order to generate the bottom loss vs grazing angle curves for different bottom types.

In order to relate the geoacoustic data in Table 3.2 with the colour map of sediment distribution, a special method was adopted here to project the geoacoustic data onto the map. First the map was a digital image which can be stored as arrays in MATLAB®. The pixels of the image determine the size of array. Secondly, the locations of model grids (latitude and longitude) were interpolated to the locations where the pixels exist. By doing so, each pixel point was assigned a geographic coordinate in latitude and longitude. Thirdly, the colour map consists of different bottom types characterised by different colours (Fig. 2.1 for instance). Each colour has a specific RGB index, making it possible to differentiate bottom types numerically. Last geoacoustic parameters for different bottom types (see Table 3.2) were assigned to the image based on the RGB index.
Table 3.2 Geoacoustic parameters for HARCAM

<table>
<thead>
<tr>
<th>Sediment type</th>
<th>Sound speed ratio</th>
<th>Density ratio</th>
<th>Attenuation of longitudinal waves (dB/m/kHz)</th>
<th>Reflection loss (dB/bounce)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Clay(^a)</td>
<td>0.994</td>
<td>1.421</td>
<td>0.2</td>
<td>13.5</td>
</tr>
<tr>
<td>Silt(^a)</td>
<td>1.057</td>
<td>1.74</td>
<td>0.8</td>
<td>12</td>
</tr>
<tr>
<td>Muddy sand(^a)</td>
<td>1.115</td>
<td>1.856</td>
<td>0.67</td>
<td>8.5</td>
</tr>
<tr>
<td>Sand(^a)</td>
<td>1.145</td>
<td>1.941</td>
<td>0.52</td>
<td>8.0</td>
</tr>
<tr>
<td>Gravelly sand(^a)</td>
<td>1.201</td>
<td>2.034</td>
<td>0.46</td>
<td>7.5</td>
</tr>
<tr>
<td>Sandy Gravel(^b)</td>
<td>1.250</td>
<td>2.1</td>
<td>0.4</td>
<td>7</td>
</tr>
<tr>
<td>Chalk(^b)</td>
<td>1.6</td>
<td>2.2</td>
<td>0.2</td>
<td>5</td>
</tr>
<tr>
<td>Limestone(^b)</td>
<td>2.0</td>
<td>2.4</td>
<td>0.1</td>
<td>3.5</td>
</tr>
</tbody>
</table>

\(^a\)Hamilton et al., 1982; \(^b\)Jensen et al., 2011.

3.2.4 Configuration of HARCAM

HARCAM is designed to calculate TL with maximum flexibility for input parameters. All input and output files are prescribed text files. Two input files are required to perform TL calculations over a two-dimensional transect: (i) the input environmental file and (ii) the control file. The former consists of range-dependent temperature, salinity, water depths and sediments data, which are specified for each water column profile. The latter must be accompanied with the environmental file, containing the model setup. Approaches to produce these two files are described in the next two sections.

3.2.4.1 The input environmental file

A typical waveguide for a well-mixed water column is shown in Fig. 3.4, showing a three-layer structure: the sea surface, the water column and the sediment. The source deployed in the water medium generates a number of rays through different directions, propagating to the boundaries where reflection and refraction occur. The propagation of rays in the water is determined by the sound speed profile, hence the temperature, salinity and the water depth.
Consequently, these three properties over a water column are firstly extracted from the ocean model for HARCAM. For a single water column profile, the corresponding upper boundary (the sea surface) and lower boundary (the seabed) data are required to calculate the sea surface loss and bottom absorption. The former is calculated based on the sea surface wind speed, which was taken from the NCEP-II data set (NCEP-DOE Reanalysis-II, 2013). The latter are determined by the properties of sediments which are listed in Table 3.2. The water properties plus the upper and lower boundaries are a typical representation for a range-independent waveguide (see Fig. 3.4). The range-dependent representation of the environment for a transect is the sequentially combination of such single profile. For this study, the resolution for the environmental profiles is 2km, which is subject to the resolution of the ocean model.

![Schematic diagram showing a typical waveguide in well mixed waters](image)

**Fig. 3.4** Schematic diagram showing a typical waveguide in well mixed waters

### 3.2.4.2 The control file

The control file is compulsory for TL calculations in HARCAM, which is used to define the geometry of the waveguide, the source properties (e.g. frequency, source depth and beam angle), the grids of model domain and a set of functions for the optimisation of TL
calculations. An example of the control file is shown in Table 3.3. The vertical grid of the model domain is separated by 1 m and 20 m in the horizontal to ensure fine resolution.

The ray theory model is computationally efficient and more suitable for higher frequencies while the parabolic equation model is more practical for low frequencies (see section 3.2.2). Given that the HOGDSON model (ray theory) has been verified formally for frequencies between 150Hz and 10kHz, in this study HOGDSON is used for calculations with frequencies greater than or equal to 150Hz whilst simulations with frequencies less than 150Hz are performed using RAM (parabolic equation). Such combination is not only computationally efficient, but also ensures accurate TL calculations.
### Table 3.3 Parameters for HARCAM simulation

<table>
<thead>
<tr>
<th>Variable name</th>
<th>Value</th>
<th>Limit</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>Environmental file name</td>
<td>*.txt</td>
<td>N/A</td>
<td>The name of environmental file</td>
</tr>
<tr>
<td>Output file name</td>
<td>*.txt</td>
<td>N/A</td>
<td>The name of output file</td>
</tr>
<tr>
<td>Max range</td>
<td>40km</td>
<td>&gt;0.01</td>
<td>The maximum range of propagation</td>
</tr>
<tr>
<td>Source depth</td>
<td>7m</td>
<td>&gt;=1</td>
<td>Specify the depth of source</td>
</tr>
<tr>
<td>Max ray angle</td>
<td>+80°C</td>
<td>±88</td>
<td>The upper ray angle</td>
</tr>
<tr>
<td>Max ray angle</td>
<td>-80°C</td>
<td>±88</td>
<td>The lower ray angle</td>
</tr>
<tr>
<td>Ray angle step</td>
<td>0.01</td>
<td>&gt;0.00001</td>
<td>The resolution of ray angle</td>
</tr>
<tr>
<td>Frequency</td>
<td>1000Hz</td>
<td>&gt;=1</td>
<td>Specify the frequency of source</td>
</tr>
<tr>
<td>Wind</td>
<td>18KTS</td>
<td>&gt;0</td>
<td>Wind speed used to calculate surface loss</td>
</tr>
<tr>
<td>Number of receivers</td>
<td>100</td>
<td>2-500</td>
<td>Specify the number of receiver in vertical</td>
</tr>
<tr>
<td>Vertical spacing of receivers</td>
<td>1m</td>
<td>&gt;=1</td>
<td>Specify the vertical resolution</td>
</tr>
<tr>
<td>Range step</td>
<td>20m</td>
<td>&gt;1</td>
<td>Specify the horizontal resolution</td>
</tr>
<tr>
<td>PH</td>
<td>8.1</td>
<td>&gt;0</td>
<td>PH value used to calculate absorption loss</td>
</tr>
<tr>
<td>Bottom type</td>
<td>2(Sand)</td>
<td></td>
<td>Used to calculate seabed reverberation</td>
</tr>
<tr>
<td>Shadow mode</td>
<td>on</td>
<td>On/Off</td>
<td>Optimise the TL calculations at shallow zones</td>
</tr>
<tr>
<td>Spike Filter</td>
<td>on</td>
<td>On/Off</td>
<td>Prevent the occurrence of excessive spikes</td>
</tr>
<tr>
<td>Run mode</td>
<td>Incoherent/Incoherent</td>
<td></td>
<td>Select run model</td>
</tr>
<tr>
<td>Beam application</td>
<td>OMNI</td>
<td></td>
<td>Define the bean pattern of source</td>
</tr>
<tr>
<td>Smooth profiles below 500Hz</td>
<td>Off</td>
<td>On/Off</td>
<td>Optimise the diffraction leakage</td>
</tr>
<tr>
<td>SMH surface loss</td>
<td>On</td>
<td>On/Off</td>
<td>Use SMH sea surface loss algorithms</td>
</tr>
<tr>
<td>Generate ray trace data</td>
<td>On</td>
<td>On/Off</td>
<td>Output ray trace data</td>
</tr>
</tbody>
</table>
3.3 Basic statistical methods

This section introduces the basic statistical methods used for data processing, including the bias of mean (BoM), the root mean square error (RMSe), the correlation coefficient (R) and the Willmott skill parameter (Willmott, 1981).

Consider two variables $M_n$ and $O_n$, (e.g. obtained from a model and observations) which are defined at the same $N$ discrete points either in space or time. The BoM measures the average magnitude of differences between the two variables and is given by

\[ BoM = \frac{\sum_{n=1}^{N} (M_n - O_n)}{N} \]  
(3.27)

The RMSe gives an estimate of random differences between two fields. In order to isolate the effect of BoM from RMSe, it is applied to the deviations from the mean rather than to the original values. The RMSe is defined by the equation:

\[ RMSe = \left\{ \frac{1}{N} \sum_{n=1}^{N} \left[ (M_n - \bar{M}) - (O_n - \bar{O}) \right]^2 \right\}^{1/2} \]  
(3.28)

where $\bar{M}$ and $\bar{O}$ are the means of two patterns over the discrete points.

The correlation coefficient quantifies the strength and direction of the linear relationship between two variables and is defined as

\[ R = \frac{1}{\sigma_m \sigma_o} \left( \frac{1}{N} \sum_{n=1}^{N} (M_n - \bar{M})(O_n - \bar{O}) \right) \]  
(3.29)

where $\sigma_m$ and $\sigma_o$ are the standard deviations of the two variables.

The Willmott skill parameter is another measure which is used to assess how close two sets of data are. It is defined as
\[ W = 1 - \frac{\sum |M_n - Q_n|^2}{\sum (|M_n - \bar{M}| + |Q_n - \bar{Q}|)^2} \]  \hspace{1cm} (3.30)

The value of Skill=1 represents a perfect match while Skill=0 means no match.
Chapter 4

4 Ocean modelling

In this chapter, tidal and full model validations are performed by comparison to observational data. Subsequently, the sensitivity of modelled results to different atmospheric forcing is carried out in order to improve the predictability of POLCOMS in the Celtic Sea.

4.1 POLCOMS validation

4.1.1 Tidal validation

For the simulation of tidal validation, the model was run with real bathymetry, constant temperature and salinity and tidal open boundary forcing for the month of July 1998. As introduced in section 3.1.2, a total of 11 tidal constituents extracted from the ES2008 model were used to condition POLCOMS. The validation data, both the $u$ and $v$ velocities, were also obtained from the ES2008 model as it has been validated against numerous tidal gauges with the elevation RMSe less than ~2cm (Egbert et al., 2010). As the comparison was based on the statistical calculations of hourly tidal velocity fields over one month, a full comparison over all model grids is impractical for this case due to a large dataset, which exceeds the memory limit of the ES2008 model. Instead, as shown in Fig 4.1, 23 points, distributed over the whole model domain, were selected to validate the tidal velocities. This is sufficient for statistical calculations. To neglecting the effect of the bottom friction, the surface velocities of the model were used for the validation. For each point, both $u$ and $v$ components were compared with ES2008 tidal model over the time series in July 1998.
Three examples of comparison points (point 12, 14 and 20) are shown in Fig. 4.2, 4.3 and 4.4 respectively. Point 12 is located in the seasonally stratified region with relatively deep bathymetry. It can be seen from Fig 4.2 that the semi-diurnal signal with a frequency of ~12 hours is evident since the semi-diurnal $M_2$ and $S_2$ are dominant tidal components in the Celtic Sea (Pingree, 1980). The $u$ and $v$ components modelled by POLCOMS agree very well with that of ES2008 as shown in the figure. Two neap-spring circles are predicted with maximum stream ~0.36ms$^{-1}$ at this location which is comparable to a previous study (e.g. Huntley, 1980). Point 14 (Fig. 4.3) is selected since it is located in the approximate area where summer tidal fronts exist. The $u$ component is well modelled while the $v$ component is underestimated slightly. The resultant velocity at spring tides is higher than that of point 12 by ~0.16ms$^{-1}$ due to shallower bathymetry. Point 20 (Fig. 4.4) positioned near the Bristol Channel is known as
high tides due to its shallow bathymetry. The magnitude of tides can reach 1.02ms\(^{-1}\) when tides propagate to such shallow regions as shown in the figure. The \(u\) component matches well with ES2008, but the \(v\) component is underestimated by 0.1ms\(^{-1}\). This is most likely caused by the different sources of bathymetry used by POLCOMS and ES2008.

Tidal modelling has been, to a large extent, a robust technology as a component of ocean models. The accuracy of barotropic tidal predictions in the European shelf seas has been investigated in a number of studies (e.g. Davies and Jones 1992; Kwong et al., 1997; Holt et al., 2005). A statistical summary for the tidal comparison of this model is shown in Table 4.1. For the majority of areas the model can reasonably reproduce the tides with low mean errors and RMSe. The model appears to underestimate the overall tides since the majority of points show negative values. The errors are comparable to previous studies (e.g. Kwong et al., 1997; Holt et al., 2005) which are summarised in Table 4.1, giving an equivalent accuracy statistically. The errors of this model are calculated from 11 constituents over the 23 points, with the overall mean errors being 0.476cms\(^{-1}\) and 0.388cms\(^{-1}\) for \(u\) and \(v\) component respectively, while the mean errors of \(M_2\) and \(S_2\) components predicted by a ~7km model of the European continental shelf waters are 0.8cms\(^{-1}\) and 0.6cms\(^{-1}\) (Holt et al., 2005). The RMSe of this model (~2km) does not show improvements compared with either the ~7km model of Holt et al. (2005) or a ~12km model from Kwong et al. (1997). The mean errors of this model are improved whilst the RMSe are typical as previous studies (see Table 4.1). This suggests that the model is sufficient to predict the overall tides over this region. However, in extremely shallow waters and sharp bathymetry regions (e.g. point 5, 19 and 23) the errors are increased (see Table 4.1). This is likely due to the inaccurate representation of bathymetry, which is of great importance to the prediction of tidal currents as discussed by Jones and Davies (1996).
Fig. 4.2 Comparison of $u$ (upper) and $v$ (lower) tidal components between ES2008 model and POLCOMS output at point 12 as shown in Fig. 4.1.
Fig. 4.3 Comparison of $u$ (upper) and $v$ (lower) tidal components between ES2008 model and POLCOMS output at point 14 as shown in Fig. 4.1.
CHAPTER 4. OCEAN MODELLING

Fig. 4. 4 Comparison of $u$ (upper) and $v$ (lower) tidal components between ES2008 model and POLCOMS output at point 20 as shown in Fig. 4.1.
Table 4.1 Statistical summary of tidal validation (POLCOMS minus ES2008). RMSe expresses the root-mean-square error.

<table>
<thead>
<tr>
<th>Points</th>
<th>Mean error: ( u ) (cms(^{-1} ))</th>
<th>Mean error: ( v ) (cms(^{-1} ))</th>
<th>RMSe: ( u ) (cms(^{-1} ))</th>
<th>RMSe: ( v ) (cms(^{-1} ))</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.112</td>
<td>-0.164</td>
<td>5.649</td>
<td>4.304</td>
</tr>
<tr>
<td>2</td>
<td>0.498</td>
<td>0.319</td>
<td>5.137</td>
<td>4.210</td>
</tr>
<tr>
<td>3</td>
<td>0.349</td>
<td>-0.457</td>
<td>5.478</td>
<td>3.949</td>
</tr>
<tr>
<td>4</td>
<td>-0.157</td>
<td>0.857</td>
<td>6.126</td>
<td>5.851</td>
</tr>
<tr>
<td>5</td>
<td>-0.082</td>
<td>-0.106</td>
<td>7.554</td>
<td>10.973</td>
</tr>
<tr>
<td>6</td>
<td>-0.583</td>
<td>-0.187</td>
<td>5.066</td>
<td>2.152</td>
</tr>
<tr>
<td>7</td>
<td>-0.310</td>
<td>-0.476</td>
<td>5.015</td>
<td>1.864</td>
</tr>
<tr>
<td>8</td>
<td>-0.323</td>
<td>-0.712</td>
<td>5.954</td>
<td>1.732</td>
</tr>
<tr>
<td>9</td>
<td>-0.222</td>
<td>-0.469</td>
<td>7.333</td>
<td>1.340</td>
</tr>
<tr>
<td>10</td>
<td>0.277</td>
<td>1.002</td>
<td>8.148</td>
<td>4.743</td>
</tr>
<tr>
<td>11</td>
<td>-0.732</td>
<td>-0.211</td>
<td>3.989</td>
<td>0.881</td>
</tr>
<tr>
<td>12</td>
<td>-0.650</td>
<td>-0.925</td>
<td>4.078</td>
<td>0.683</td>
</tr>
<tr>
<td>13</td>
<td>-0.810</td>
<td>-1.295</td>
<td>4.247</td>
<td>0.919</td>
</tr>
<tr>
<td>14</td>
<td>-0.496</td>
<td>-0.317</td>
<td>6.324</td>
<td>1.833</td>
</tr>
<tr>
<td>15</td>
<td>-0.306</td>
<td>0.020</td>
<td>8.575</td>
<td>5.251</td>
</tr>
<tr>
<td>16</td>
<td>-0.726</td>
<td>-0.048</td>
<td>2.463</td>
<td>0.509</td>
</tr>
<tr>
<td>17</td>
<td>-1.066</td>
<td>-0.359</td>
<td>3.822</td>
<td>0.351</td>
</tr>
<tr>
<td>18</td>
<td>-1.088</td>
<td>-1.878</td>
<td>3.368</td>
<td>6.643</td>
</tr>
<tr>
<td>19</td>
<td>-0.677</td>
<td>-0.532</td>
<td>10.837</td>
<td>6.642</td>
</tr>
<tr>
<td>20</td>
<td>-0.654</td>
<td>-0.936</td>
<td>4.432</td>
<td>2.238</td>
</tr>
<tr>
<td>21</td>
<td>-0.455</td>
<td>-0.109</td>
<td>4.630</td>
<td>0.759</td>
</tr>
<tr>
<td>22</td>
<td>-1.382</td>
<td>-0.231</td>
<td>7.906</td>
<td>0.522</td>
</tr>
<tr>
<td>23</td>
<td>-1.476</td>
<td>-1.726</td>
<td>9.411</td>
<td>4.913</td>
</tr>
</tbody>
</table>

**Total of this model**

| M\(2\) of Holt et al. (2005) | 0.8 | 5.8 |
| S\(2\) of Holt et al. (2005) | 0.6 | 2.5 |
| M\(2\) of Kwong et al. (1997) | -- | 5.7 |
| S\(2\) of Kwong et al. (1997) | -- | 2.2 |
4.1.2 Full model validation

4.1.2.1 Introduction

As the barotropic tides have been validated by comparison with the ES2008 model, a full model validation is indispensable to assess the overall model performance. The year 1998 was chosen to be modelled since there were two fieldwork campaigns in July and August 1998, giving extensive Scanfish data for validations. The simulation started from 1st June 1998 and ended on 15th September 1998, with barotropic and baroclinic time steps of 2s and 20s respectively. Such time steps have been found to be fine enough to predict the currents pattern and kinetic energy variations in the Celtic Sea for a 2km model (Shapiro, 2011). The model was forced by tides and surface forcing (NCEP-II, 1.6° resolution) and the model set-up can be found in section 3.1.2.

Because the model output is in binary format, a set of MATLAB® scripts were developed to perform the data post-processing with following functions:

a) Extracting data from the binary file (hourly output for this study)

b) Interpolate data from s levels to z levels

c) Save both s-level data and z-level data in MATLAB® format which is easier to manipulate in subsequent processes

d) Visualization and statistical analysis

4.1.2.2 Geostrophic simulation

In order to provide the initial currents for the full model run, a geostrophic run was performed, as it allows the currents to spin up without losing the initial temperature field. Simulations were initialised using the “semi-diagnostic adjustment method”, i.e. by running the model with no meteorological forcing and river discharges with frozen initial temperature and
salinity distributions (Enriquez et al., 2005). This allows the currents to achieve equilibrium with the density field. The output binary file produced by the geostrophic run was then given to the full model simulation so that the full simulation could be initialised from a warm start.

The geostrophic simulation was run from the 1st of June to the 7th of June 1998, with no tidal and meteorological forcing. This time length was selected because after seven days (168 hours) the volume averaged kinetic energy over the whole basin reaches stability, thus preventing or minimising spurious currents. The time length is always chosen to be sufficiently long to achieve a state of intermediate asymptote with a nearly constant energy level, but short enough not to introduce significant modification to the initial temperature and salinity fields (Enriquez et al., 2005).

![Volume averaged kinetic energy for the geostrophic run](image)

**Fig. 4.5** Volume averaged kinetic energy for the geostrophic run

Fig. 4.5 shows the volume averaged kinetic energy (KE) for the geostrophic run over the whole month of June 1998. KE increases rapidly at the beginning of the simulation with oscillations, reaching a constant level of \(~0.15\ \text{Jm}^{-3}\) after approximately 160 hours. After that
the amplitude of oscillations decays gradually, maintaining a near constant level. A seven-day simulation (168 hours) is, therefore, sufficient to adjust the initial currents fields for this study.

4.1.2.3 Observational data

The temperature and salinity structure of the Celtic Sea was examined by Brown et al. (2003) using a towed undulating CTD (Scanfish) on 19 transects in the Celtic Sea between 26 August and 5 September 1998 (hereafter referred to as Cor98). Four of these transects (numbered 182, 202, 189 and 187), which are located within the model domain as shown in Fig. 4.6, have been used for the validation. The data cover the depth range between 4m below the sea surface and 5m above the seabed and have a 1m vertical resolution and a horizontal separation of approximately 150-500m.

![Fig. 4.6](image.png) Locations of four Scanfish sections selected to validate the model, indicated as the black solid lines and numbered as 182, 202, 189 and 187 respectively. The colour scale shows bathymetry in meters.
Due to the difference in resolution between the Cor98 data and the POLCOMS model (1-4m in the vertical by 2km in the horizontal), a common grid with an intermediate resolution (2m in the vertical by 1km in the horizontal) was used to carry out comparisons. First the original Cor98 data were interpolated onto a common grid along each transect. Interpolation was achieved using the Krigeing method with a search radius of 1.5km in the horizontal and 5m in the vertical and then interpolating the 3D hourly POLCOMS outputs onto the same grid. As the Scanfish transects took several hours to complete, they have been split into 1-hour segments to match the model output for the same hour. Finally, a synthetic model output from individual 1-hour blocks was generated to cover the whole period of the transects and compare the model output to the observations.

4.1.2.4 Assessment of model performance

The performance of the model has been evaluated by comparison of model results with observed Scanfish data collected using towed undulating CTD on four sections between August and September 1998. The comparisons focus on the thermal structures because the water column structure is dominated by the temperature in most of regions in the Celtic Sea (Brown et al., 2003). The main features of the transects and their comparison with model outputs are discussed below:

Section 182:

Section 182 (Fig. 4.6) crosses the model domain from the model western boundary to the Cornish coast over the Celtic deep, capturing the main thermal features of the Celtic Sea. The comparison of the temperature structure between Cor98 Scanfish data and POLCOMS is shown in Fig. 4.7. The top panel shows the temperature structure observed on 27 August in 1998 while the bottom one illustrates the model results. The model reproduces the complex thermal features observed, predicting the intense thermocline through the centre to the south.
Fig. 4.7 Comparison of temperature structure along section 182 between Cor98 Scanfish data (top panel) and model outputs (bottom panel).
Furthermore, small variations along the thermocline due to the interaction between wind induced mixing and bottom tidal mixing are modelled accurately, giving a good agreement here.

The cold bottom dome (<11°C) over the Celtic deep (distance 95-160km from the starting point of the transect) is resolved clearly, although with slightly colder water (~0.2°C) than observed. A secondary dome expanding from 160km to 200km associated with the bottom front is also reproduced successfully by this model, but with less horizontal diffusion in the modelled results. The modelled surface temperature shows a peak of ~17°C ranging from 100km to 220km which is similar to observations, although it is approximately 0.5°C colder near the Cornish coast (230-250km), most likely resulting from the insufficient surface heat flux. However, the north-western margin of the cold bottom dome (close to the open boundary, left of 100km mark in Fig. 4.7) is predicted without showing the front below 50m which is evident in observations. As the tides within this model were validated carefully, it is unlikely that this discrepancy is due to insufficient magnitude of tidal currents. A potential reason for reduced accuracy near the open boundary of the domain is the poor resolution of boundary temperature and salinity data extracted from 0.25° monthly climatology of the World Ocean Atlas (Boyer et al., 2004). Fig. 4.7 shows that at the western boundary water column is much colder in August than observed, particularly below a depth of 30m. This cold water is advected horizontally from the open boundary thus hindering formation of the front by tidal mixing near the 80km mark in Fig. 4.7.

In order to assess quantitatively the model skill, section 182 is divided into four layers in the vertical, labelled as A (Surface mixed layer), B (Stratified layer), C (Intermediate layer) and D (Bottom layer) respectively. The statistical analysis is summarised in Table 4.2, which also shows the depth for each layer and the number of points sampled for statistical calculations.
Table 4.2 shows a very good model skill in the surface mixed layer A, with BoM = -0.02 °C (model is slightly colder); and RMSe less than 0.4 °C. Despite missing one of the temperature fronts in the bottom layer, the model still performs reasonably well there with slightly higher RMSe of 0.92°C. Although the thermocline location and variations are well reproduced, given the complexity of the temperature pattern in the stratified layer, the temperature difference in layer B is higher (RMSe = 1.26°C) than in other layers. In terms of model performance, the model demonstrates good accuracy in modelling thermal structures of the Celtic sea, giving high values for both correlation coefficient (R=0.77 to 0.92) and the Willmott skill parameter (W=0.70 to 0.93). The values of BoM, RMSe, correlation coefficient and model skill for the transect as whole are BoM= -0.72°C, RMSe=0.99°C and W=0.93, which is similar to the performance of the best operational models.

**Table 4.2 Statistical summary for temperature in each layer of Section 182 (BoM: bias of mean; RMSe: root mean square error; R: correlation coefficient; W: Willmott skill parameter).**

<table>
<thead>
<tr>
<th>Layers</th>
<th>Depth (m)</th>
<th>BoM (°C)</th>
<th>RMSe °C</th>
<th>R (0-1)</th>
<th>S (0-1)</th>
<th>Data points sampled</th>
</tr>
</thead>
<tbody>
<tr>
<td>A: Surface mixed layer</td>
<td>0-17</td>
<td>-0.016</td>
<td>0.39</td>
<td>0.79</td>
<td>0.89</td>
<td>2040</td>
</tr>
<tr>
<td>B: Stratified layer</td>
<td>17-40</td>
<td>-0.34</td>
<td>1.26</td>
<td>0.76</td>
<td>0.82</td>
<td>2244</td>
</tr>
<tr>
<td>C: Intermediate layer</td>
<td>40-70</td>
<td>-1.4</td>
<td>0.92</td>
<td>0.77</td>
<td>0.71</td>
<td>2610</td>
</tr>
<tr>
<td>D: Bottom layer</td>
<td>70-110</td>
<td>-0.79</td>
<td>0.54</td>
<td>0.86</td>
<td>0.70</td>
<td>1133</td>
</tr>
<tr>
<td>Whole section</td>
<td>0-110</td>
<td>-0.72</td>
<td>0.99</td>
<td>0.92</td>
<td>0.93</td>
<td>8024</td>
</tr>
</tbody>
</table>

**Section: 202**

Section 202 (Fig. 4.6) is located to the northwest of the model domain, along the axis of the Celtic deep. The comparison of thermal structure between the model and observations is shown in Fig. 4.8. According to both the model and observations, the Celtic deep is strongly stratified with an intense thermocline at a depth of approximately 30m. The temperature structure along this section is well resolved by the model, predicting the precise location of
the thermocline. In the surface mixed layer, the model underestimates temperature by ~0.5°C, presumably due to insufficient surface heat flux. Two bottom cold pools (at 0-25km and 30-54km marks) are also reproduced, although the modelled water is slightly colder (~0.3°C) than observed. Such discrepancies are probably due to the errors in the initial and boundary conditions which were taken from relatively coarse climate data. As examined by Wakelin et al. (2009), the effect of initial temperature and salinity conditions on the model outputs can be neglected after 15 months of model time, and then the meteorological forcing becomes dominant. For this study the model outputs are taken from the third month of the model time, so that the effect of initial field is reduced but cannot be ignored completely. The effect of boundary conditions is, however, felt at the edges of model domain independently to the length of simulation. At the north-eastern end of this transect, the thermocline jumps suddenly from a depth of ~33m to ~28m, which is not consistent with the observations. This is likely to be due to the effect of errors in the open boundary condition. A statistical summary for this transect is given in Table 4.3.
Section 189

Section 189 (Fig. 4.6) reveals a strong thermocline at a depth ~30m, associated with a series of bottom fronts as shown in Fig. 4.9. The intense stratification predicted is similar to the observations. At a distance of 0-20km the bottom cold water is reproduced accurately by the
model, as well as the associated bottom fronts. They are, however, more diffuse in comparison with the observations. The location of the coastal front at 100km is also modelled well, although it is slightly displaced and diffused when compared to observations. A possible reason for this is excessive tidal mixing at 100-200km, resulting from the inaccurate representation of bathymetry by ETOPO2 data used in the model.

The selection of values of horizontal diffusion (equation A1.19 in Appendix 1) is important to the accuracy of model simulations; there is, however, no sufficient theoretical basis for its choice (Holt and James, 2006). A proper horizontal diffusion coefficient can not only improve the parameterisation of unresolved eddies in a subgrid scale, but also preserve the eddy features (Holt and James, 2006; Shapiro et al., 2013). As can be seen in Fig 4.9, stronger horizontal mixing at the boundary of the bottom front (~22km) is modelled compared to the observation. For this simulation the model was run with no horizontal diffusion ($\alpha = 0$, see equation A1.19 in Appendix 1), assuming that there is sufficient numerical diffusion to account for it (Holt et al., 2005). However, the numerical diffusion can even over mark such effect as shown in Fig. 4.9, even with the use of an advection scheme (the Piecewise Parabolic Method). It is then unsurprising that including the physical diffusion will create much stronger diffusion and degrade the frontal contrast (not shown here). Consequently, the coefficient $\alpha = 0$ is considered to be the idealist one for this study since it conserves the frontal structures which are sensitive to sound propagation.

Section: 187

Section 187 (Fig. 4.6) is positioned to the north of section 189, crossing the margin of the Celtic deep eastwards to the Bristol Channel. As shown in Fig. 4.10, the model provides a temperature distribution close to that observed, with a strong stratification at a depth of ~35m and a well-mixed water column near the Bristol Channel. However, the bottom front in the
model is not far enough eastwards. As we do not take into account the river input near the Bristol Channel, the resulting lack of freshwater input in this region leads to insufficient buoyancy inputs, thus contributing to a less accurate location of the bottom front compared to other transects.

The statistical properties of sections 202, 187 and 189 are summarised in Table 4.3. The data points for statistical calculations are sampled from the entire section. The thermal structures of the transect are well resolved with high correlation coefficients and low errors (biases) in the mean values. The values of the demeaned RMSe between the model and observation on all three sections are slightly higher but still not exceeding 1.0°C. The model skill is as high as W=0.87-0.98, and the errors at the end of the transects are caused mainly by uncertainties in the boundary temperature conditions. Overall, the data shown in Table 4.3 confirm the high skill of the model in representing the temperature structure in the Celtic Sea, both at the surface and within the water column.

**Table 4.3** Statistical summary of temperature of section 202, 187 and 189 (BoM: bias of mean; RMSe: root mean square error; R: correlation coefficient; W: Willmott skill parameter).

<table>
<thead>
<tr>
<th>Sections</th>
<th>RMSe (°C)</th>
<th>BoM (°C)</th>
<th>R (0-1)</th>
<th>S (0-1)</th>
<th>Data points sampled</th>
</tr>
</thead>
<tbody>
<tr>
<td>202</td>
<td>0.53</td>
<td>-0.35</td>
<td>0.98</td>
<td>0.98</td>
<td>2795</td>
</tr>
<tr>
<td>187</td>
<td>0.81</td>
<td>-0.077</td>
<td>0.81</td>
<td>0.87</td>
<td>2886</td>
</tr>
<tr>
<td>189</td>
<td>0.83</td>
<td>-0.17</td>
<td>0.95</td>
<td>0.97</td>
<td>4335</td>
</tr>
</tbody>
</table>
Fig. 4. 9 Comparison of temperature structure along section 189 between Cor98 Scanfish data (left) and model outputs (right).
Fig. 4. 10 Comparison of temperature structure along section 187 between Cor98 Scanfish data (top panel) and model outputs (bottom panel).
4.1.3 Conclusion

The tidal currents were compared with the European Shelf Tide Model developed by Oregon State University. The statistical errors analysis gives an equivalent accuracy to previous studies and slight improvements in some regions compared with previous studies, with the typical errors in mean being -0.476cms\(^{-1}\) and -0.388cms\(^{-1}\) for \(u\) and \(v\) component respectively. This suggests that the model is sufficient to predict the tidal currents in the Celtic Sea.

The ability of the optimally configured POLCOMS model to resolve vertical thermal structures in the Celtic Sea was assessed by comparisons of model simulations with an extensive set of observational data (4 Scanfish sections selected for this study) collected in the Celtic Sea in 1998. The analysis has revealed a good level of model skill, with the mean deviations of modelled and observed temperatures of -0.72°C, -0.35°C, -0.08°C, and -0.17°C for the transects 182, 202, 189 and 187 respectively. The Willmott model skill parameter on the same transects is as high as 0.93, 0.98, 0.87 and 0.97 respectively.

Accurate simulations for the thermocline and frontal positions are indispensable for later acoustic modelling since the underwater sound propagation is of particular relevance to such features. In this study, the depth of the thermocline and the location of the bottom fronts are a primary consideration, placing much greater emphasis on the accuracy of modelled vertical structures. The positions of the thermocline and bottom fronts were well predicted, although more diffusion occurs at the boundaries of bottom fronts. This may be attributable to the numerical diffusion as the horizontal diffusion terms are switched off for the momentum and transport equations in this simulation. The results also demonstrated good capability in reproducing the sharp interface of the thermocline, which has high importance in determining the strength of reflection and refraction of sound waves.
4.2 Sensitivity study: I

In this section it examines if the performance of an ocean model is improved when the data to force the model is taken from a higher resolution atmospheric model. The models used are identical except for their meteorological forcing: the low resolution forcing was obtained from the National Centers for Environmental Prediction (NCEP-II, 1.6° resolution), and the high resolution forcing obtained from the British Atmospheric Data Centre (BADC, 0.11° resolution). Note that the BADC high resolution data are available since 2006. Both models were run for the year 2008, and comparisons were made with the SST obtained from NASA POET (http://poet.jpl.nasa.gov). Both datasets have the same frequency of 6-hour.

4.2.1 Variations in meteorological forcing

A statistical inter-comparison has been carried out to identify differences between the BADC and NCEP meteorological forcing. In all statistical comparisons, the data from NCEP have been considered as the reference field, i.e. the O-variable in equations (3.24) - (3.27). The upper panel in Fig. 4.11 shows air temperature at 2 m above sea level on 1 August 2008 from NCEP and BADC. The lower panel in the same figure shows a time series of statistical measures of the difference between the two fields. The data for calculation of the statistics are taken over the whole model domain (i.e. excluding land points) for each 6-hour time point. The spatial patterns on the upper panel are very different, and the high resolution BADC data captures clearly the differences between land and sea areas. The BADC data are warmer in winter, giving a time averaged BoM= 0.51°C, and significantly cooler in spring with an average BoM= -0.81°C, they are slightly cooler in the summer (negative BoM) giving a typical difference of -0.3°C to -0.4°C. The discrepancies measured by demeaned RMSe over the whole year are small with a typical value of 0.25-0.4°C. However the correlation
coefficient is quite low R=0.4 to 0.5 indicating dissimilarity in the spatial pattern of air temperature.

![Temperature Maps](image)

**Fig. 4.11** Sample air temperatures at 2m from BADC (upper left), NCEP (upper right) and the statistical differences for the year 2008 (bottom panel).

The difference in the cloud cover between NCEP and BADC, as shown on the upper panel in Fig. 4.12, is particularly strong. The high resolution BADC data capture small-scale variations, both over the sea and the land. The cloudiness from BADC data is consistently about 20% higher than from NCEP. The demeaned RMSe is relatively small at about 10% but a very low correlation coefficient (R=0.2-0.3) indicates a poor match between the spatial patterns of clouds provided by NCEP and BADC.
As shown in Fig. 4.13, the direction of wind speed is quite similar between BADC and NCEP whereas the magnitude exhibits difference locally. In the south of the model domain the amplitude of wind from BADC is higher than that of NCEP by ~2m/s. Overall the BoM is small with BADC slightly lower than NCEP while the RMSe maintains a near constant level (~1m/s). The spatial pattern, compared with the total cloud cover (Fig. 4.12), is comparably better with a typical higher correlation coefficient of ~0.5 (Fig. 4.13).
Fig. 4.13 Sample wind speed (m/s) from BADC (upper left), NCEP (upper right) and the statistical differences for the year 2008 (lower).

The statistics covering all four seasons of the year (winter: December, January and February; spring: March, April and May; summer: June, July and August; and autumn: September, October and November) and all 7 atmospheric variables used by the ocean model are summarised in Table 4.4. The BoM, demeaned RMSe and R were calculated using daily NCEP and BADC data. The mean sea level pressures from both data sets are very similar with extremely low bias (0.12 to 0.55 hPa) and RMSe (0.44 to 0.63 hPa), and a high correlation coefficient (R=0.73 to 0.83). Although the relative humidity and precipitation have low biases and RMSe, the spatial patterns are quite dissimilar with low correlation.
coefficients. The wind speed of BADC is consistently lower than that of NCEP, and spatial patterns are different slightly as shown in Fig. 4.12.

<table>
<thead>
<tr>
<th>Table 4.4 Summary of differences between BADC and NCEP meteorological data (BoM: bias of mean; RMSe: root mean square error; R: correlation coefficient).</th>
</tr>
</thead>
<tbody>
<tr>
<td>Winter</td>
</tr>
<tr>
<td>BoM</td>
</tr>
<tr>
<td>---</td>
</tr>
<tr>
<td>Air temperature (°C)</td>
</tr>
<tr>
<td>Total cloud cover (%)</td>
</tr>
<tr>
<td>Pressure (Pa)</td>
</tr>
<tr>
<td>Relative humidity (%)</td>
</tr>
<tr>
<td>Precipitation (kg/m²/day)</td>
</tr>
<tr>
<td>U-wind (ms⁻¹)</td>
</tr>
<tr>
<td>V-wind (ms⁻¹)</td>
</tr>
</tbody>
</table>

4.2.2 Variations in the modelled SST

The 8-day averaged SST charts simulated by the two versions of the POLCOMS model forced by BADC and NCEP data for the year 2008 have been compared with the remote sensed data obtained from NASA POET (http://poet.jpl.nasa.gov). Compared to the daily SST data, the 8-day satellite data have a better spatial coverage due to cloud filtering. The simulated and observed 8-day averaged SST are presented in Fig. 4.14a, whilst the corresponding statistical comparisons are shown in Fig. 4.14b (NCEP minus POET) and Fig. 4.14c (BADC minus POET). The time averaged values for four seasons of 2008 are summarised in Table 4.5.

The difference between SST produced with BADC forcing and POET data increases consistently from late spring, reaching the peak point at the end of August, whereas the SST from the model forced by NCEP data gives a much better match with observations. The main reason for the difference between simulated SST obtained with BADC and NCEP data is that
these data sets have significantly different values for the cloud cover, and hence downward solar radiation. In the summer, the SST from simulation forced by NCEP is overestimated by only 0.15°C while the model using BADC predicts much colder SST (-0.61°C). The discrepancy reaches a value of BoM as high as (-1.15°C) in late August.

The length scales of most energetic dynamic features in the atmosphere are defined by the Rossby radius of deformation, which is about 1000km (the typical size of a cyclone). However sub-mesoscale atmospheric patterns, such as patchiness in the cloud cover, could result in smaller scale variations of both the wind and solar radiation, hence creating a direct link between these smaller atmospheric features and the ocean mesoscale variability, which has a natural length scale of 10-20 km in a shelf sea.

Fig. 4. 14  (a) The time line of 8-day averaged sea surface temperature from models driven by BADC and NCEP forcing and remotely sensed data from NASA POET. (b) Statistical differences between SST simulated with NCEP forcing and satellite data as a function of time. (c) Statistical differences between SST simulated with BADC forcing and satellite data as a function of time.
The analysis indicates that in the Celtic sea, the SST produced by the same model is extremely sensitive to variations in meteorological forcing during the summer. The results support the conclusion by O’Neill et al. (2012) who claimed that the resolution of surface forcing has significant impact on the seasonal cycle of sea surface and bottom temperatures. However, the results show that the main reason behind the differences is not the resolution of the forcing as such, but a significant difference in the mean values of cloudiness, wind speed, and air temperature. These factors control the majority of sea-air heat exchange and hence the sea surface temperature. The total cloud cover in BADC data is consistently much higher than that of NCEP, so that the resulting solar short wave radiation is lower, which in turn leads to inadequate surface heat flux being transferred to the ocean. This effect is partly compensated by lower winds represented in BADC, which reduces evaporation and hence the loss of latent heat from the sea. Lower air temperatures in the spring and beginning of summer in the BADC data also contribute to lower SST produced by the model. In contrast to the situation in the Irish Sea reported by O’Neill et al (2012), higher resolution meteo data over the Celtic Sea do not necessarily result in better representation of the SST by the ocean model. To the contrary, it is the low resolution NCEP data which resulted in better simulation of the SST, particularly in the summer months.
The resolution of surfacing forcing is an important consideration for numerical modelling. The differences in time and space of forcing can lead to significantly dissimilar responses of oceanic processes based on modelling, e.g. the eddy kinetic energy, the mixed layer depth, the heat budget and the currents pattern (e.g. Cravatte and Menkes, 2009; Brossier, et al., 2011). There are many worldwide sources to extract meteorological forcing (e.g. NCEP and JRA), which are widely utilised to drive ocean models for various implementations. For the same database the model results the model results can be improved by increasing the spatial and temporal resolution as examined by O’Neill et al. (2012). However, this study indicates that this is due not to the resolution of the forcing per se, but to the differences between the meteorological models in mean values of parameters such as total cloud cover, which in turn reduce the solar radiation flux reaching the sea surface in the oceanographic model.

4.2.3 Variations in the modelled heat budget
Discrepancy in SST indicates different heat budget equilibrium. In order to examine the reasons and identify which parameter resulting in such difference in SST, the variability of surface heat budget is evaluated shown as below.

The total outgoing heat flux averaged over eight days in time and the model domain in space and its components are shown in Fig. 4.15. Overall, the contribution from sensible heat flux to the heat budget is mild whereas the longwave radiation and latent heat flux are comparatively larger in this region. The latent heat flux plays a significant role in determining the variability of total heat loss with stronger variability occurring during late autumn, winter and early spring.
Fig. 4.15 Sea surface outgoing heat flux (8-day averaged) containing: HL: the out-going longwave radiation; SK: the sensible heat flux; SKE: the evaporative heat flux.

Specifically, the sensible heat flux determined by wind speeds and air-sea temperature contrast exhibits small differences between BADC and NCEP. The longwave radiation from NCEP is consistently higher than that of BADC by ~5Wm$^{-2}$ over the year whilst the latent heat of NCEP is only slightly stronger in summer. This suggests that the cloudiness of BADC is denser than that of NCEP as indicated in table 4.4, leading to stronger longwave radiation loss. The total heat loss of NCEP is persistently higher over the year, with greater magnitude occurred in summer as shown in Fig. 4.15.

Fig. 4.16 Heat flux of short wave radiation (8-day averaged)
Fig. 4.16 illustrates the total heat flux gained by the sea surface for BADC and NCEP, which is generated by the shortwave radiation. A strong seasonal signature is observed since the shortwave radiation is the most influential parameter in driving the seasonality of SST in midlatitudes (Kara et al., 2009). It is evident that the heat input simulated by NCEP maintains consistently higher over the year with larger magnitude of difference between April and September, reaching a maximum of \(\sim 130 \text{ wm}^{-2}\) in July.

The net heat flux for both simulations is shown in Fig. 4.17. Although the total heat loss of NCEP (Fig. 4.15) is larger, the net heat input of NCEP is still remarkably higher (\(\sim 45 \text{ wm}^{-2}\)) between April and September. However, during the period of cooling it demonstrates mild differences, which indicates an imbalance of the annual heat budget. For example, the net heat budget over the entire year predicted from NCEP is higher than that of BADC. This potentially affects the inter-annual evaluation of SST and heat content when using different surface forcing.

![Net heat flux into the sea surface (W/m²)](image)

**Fig. 4. 17** Net heat flux (8-day averaged). Negative values indicate the heat gain and positive values mean the loss.
As seen in Fig. 4.16, the incoming solar radiation (the short wave radiation) is the main cause for such large difference in the net heat flux. The inward flux calculations formulated by the bulk formulas from Elliott and Clarke (1991), following Gill (1982), and are defined as:

\[ q_{in} = q_s C_0 \times (1 - \alpha) \]  
\[ C_0 = (1 - c \times c_c - 0.38c^2) \]

where \( q_{in} \) is the inward flux from short wave radiation (Fig. 4.16), \( q_s \) is the solar radiation available to the surface calculated from astronomical parameters, which are identical for both simulations and \( c_c \) is the cloud cover coefficient and \( \alpha \) is the sea surface albedo. It is indicated by equation 4.1 and 4.2 that the incoming flux is governed completely by the total cloud cover \( c \). Fig. 4.18 shows the variation of \( C_0 \) for NCEP and BADC. Due to much denser cloud cover from BADC (see Table 4.4) constantly lower values of \( C_0 \) are calculated over the year, correspondingly resulting in large difference in short wave radiation flux, hence SST. As \( q_{in} \) is proportional to the total radiation flux \( q_s \), the difference of incoming flux between BADC and NCEP grows along with the increase of \( q_s \) when the difference of \( C_0 \) remains a
nearly constant level. This can explain why the largest difference occurs in late July (see Fig. 4.14) where strongest solar radiation exists.

Apart from the SST, such large difference in heat flux may lead to greatly different responses of stratification and frontal structures since the Celtic Sea is stratified generally from April to November (Pingree, 1980). Differences in heat flux and wind forcing can create direct link to the stratification over the water column. This may lead to different water column structures. In terms of turbulence, the turbulence kinetic energy (TKE) is in the balance that it is created and consumed by diffusion, shear production, mixing and dissipation. For a stable stratification the TKE can be converted to potential energy through mixing or buoyancy production whilst in an unstable situation the loss of potential energy can contribute to the TKE via convection. The shear production is dependent on the vertical shear produced by the horizontal velocities while the mixing is related to the vertical density gradient. The dissipation rate can be also affected since this model uses the two-equation and dynamic dissipation rate scheme (\(k-\varepsilon\) type), where the dissipation rate is calculated utilising a structurally similar equation as the TKE (see Holt and Umlauf, 2008). Different surface inputs can lead to changes of the terms described above through various mechanisms, hence breaking the local equilibrium between TKE, shear production, mixing and dissipation. This may affect the stratification and the estimation of eddy diffusivity, hence the vertical mixing and turbulence flux.

Fig. 4.19 shows the temperature structures in summer for BADC and NCEP where strong stratification occurs. In the upper mixed layer the temperature of NCEP is higher than that of BADC by \(\sim 1.7^\circ C\) whereas at the bottom layer the difference is small. The depth of thermocline which is marked by the solid line (Fig. 4.19) also exhibits strong discrepancy. This highlights the needs to investigate the sensitivity of surfacing forcing to the stratification
parameters (e.g. the depth of thermocline and the position of bottom fronts). This work will be presented in the next section where more meteorological forcing is used.

**Fig. 4.19** Snapshots of temperature structure in summer along N° 51.36 for BADC (left) and NCEP (right). Solid lines mark the depth of thermocline.

### 4.2.4 Conclusion

The inter-comparison between high resolution BADC meteorological forcing data and coarser resolution NCEP data demonstrates strong dissimilarities, which result in differences in the SST produced by the ocean model driven by those data. The results indicate that simulated SST is less sensitive to the differences between high resolution BADC and low resolution NCEP data during the winter and spring despite significant differences in the meteorological data.

During the summer and autumn time, however, the modelled SST hindcasts fit the observations better when a coarser resolution (NCEP) meteorological forcing is used, with the model output using the high resolution BADC meteorological data being persistently lower than observations. Errors increase during the summer, reaching approximately -1.15°C in August after the period of strongest solar radiation in July. As the BADC and NCEP data vary in their mean values of cloudiness, wind speed, and air temperature, this in turn
influences sea-air heat exchange and thus the sea surface temperature output of the POLCOMS model. The total cloud cover is of the greatest importance to such modelled SST difference, showing time dependence with higher difference occurring in summer. Refinements in resolution of the meteorological forcing do not, therefore, necessarily correlate directly with improvements in oceanographic model performance and should not be considered in isolation.

The choice of meteorological forcing may also be highly influential on the seasonal stratification, the vertical mixing and maybe the bottom frontal position. A sensitivity study is essential to clarify this question and this is examined in the following section.

4.3 Sensitivity study: II

This section presents the sensitivity study of modelled SST and water column structures to different atmospheric forcing. Five re-analysis datasets extracted from different sources were used since they are popular worldwide database frequently utilised to force oceanic models due to their better spatio-temporal coverage. The responses of SST, the depth and strength of the thermocline and the position of bottom fronts to different meteorological forcing are presented in this section.

4.3.1 Introduction

Seasonal stratification and associated tidal mixing fronts are common features in the continental shelf seas. The spatial temperature structure and the variability are largely dependent on the vertical mixing processes over the water column in the Celtic Sea, which are controlled in turn by the mechanisms: potential energy and turbulent mixing (Simpson and Hunter, 1974; Holt and Umlauf, 2008). The former is governed by surface heat flux while the latter is a result of tides and winds. Different atmospheric forcing can thus
potentially result in dissimilar water column structures through vertical mixing, generated by boundary layers and internal mixing processes.

In a non-tidal basin a previous study by Shapiro et al (2011) has shown that the basin-wide circulation pattern and the temperature structure in the Black Sea produced by the same model are significantly dependent on the source of the meteorological input whereas in the tidally active seas some physical features (e.g. the frontal position) are not sensitive to the variations of surface heat flux (Holt and Umlauf, 2008) and wind forcing (Simpson et al., 1978). Although tides have potential to limit the effects of surface forcing, the wind forcing can be still a significant contributor to the total mixing (Simpson et al., 1978). A sensitivity study performed by O’Neill et al. (2012) examined the effects of varying resolution of surface forcing on the predictive skills of simulating the sea surface temperature and salinity in a tidally active sea. In accordance with their study, the RMSe of SST can be reduced by 20% - 30% by improving the spatiotemporal resolution of atmospheric forcing. Young and Holt (2007) demonstrated strong variability of the onset and breakdown of stratification on interannual scales in the Irish Sea based on a 40-year simulation, which highlights the necessity to investigate the sensitivity of water column structure to meteorological forcing in a tidally dominant basin.

Seasonal stratification and associated tidal mixing fronts are of great biological (see e.g. Pingree, 1980; Sharples et al., 2001) and acoustic (e.g. Katsnelson, et al., 2012; Shapiro et al., 2014) importance. The seasonal thermocline, separating the high-light low-nutrient surface water from the low-light high-nutrient bottom water, is often found to be the region of enhanced chlorophyll concentration (Sharples et al., 2001). The thermocline is also considered as a physical barrier, which inhibits vertical mixing and vertical diffusion (Pingree et al., 1977). Phytoplankton trapped in the surface mixed layer can receive sufficient
irradiated light and stronger photosynthetic rate than that in stratified waters due to effects of mixing. Limits on vertical diffusion can also prevent the nutrition from bottom water penetrating the thermocline to the surface layer, hence reducing the primary production in the upper layer. In terms of the acoustics, the stratification and the associated fronts can cause the difference in the level of sound energy as high as ~20dB (e.g. Heathershaw et al., 1991; Lynch et al., 2003).

The coupled ocean-biological (e.g. Holt et al., 2012) and ocean-acoustic (e.g. Lermusiaux et al., 2010) models are now popular tools used for different implementations. Consequently, investigating the sensitivity of details of thermocline and bottom fronts is beneficial to promote the accuracy of coupled model systems.

4.3.2 Variations in meteorological forcing

Five reanalysis products used in this study include the ECMWF 40-year Reanalysis (ERA40), ECMWF Interim Reanalysis (ERA-Interim), Japanese 25-year Reanalysis (JRA25), Japanese 55-year Reanalysis (JRA55) and NCEP/DOE Reanalysis II (NCEP), details of which are summarised in Table 4.6. All products have an identical time interval of 6-hour, but have different spatial resolutions.

<table>
<thead>
<tr>
<th>Table 4.6 Summary of five meteorological reanalysis products</th>
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<tr>
<td>Resolution (Lat×Long)</td>
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<tr>
<td>0.75°×0.75°</td>
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<td>Time step</td>
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The ERA40 dataset is a global atmospheric reanalysis product from 1957 to 2002, produced using the ECMWF integrated Forecast Model with a spatial resolution of ~125km and 60 vertical levels. Various observational data are assimilated using a 6 hourly 3D variational analysis. Satellites data (e.g. Vertical Temperature Profile Radiometer radiances) are incorporated into the reanalysis field and the motion winds are also used. ERA-interim is a new generation extended from REA40, provided with a higher resolution of 0.75°×0.75° in space using an improved version of the ECMWF integrated Forecast Model. The data assimilation of ERA-interim is based on a 12-hourly four-dimensional variational analysis with adaptive estimation of biases in satellite radiance data. The JRA25 product is the first long-term global atmospheric reanalysis produced in Asia. The dataset is generated using the Japan Meteorological Agency numerical assimilation and forecast system, with a resolution of 1.25°×1.25° and 40 vertical levels. The observational data used for the assimilation system are collected specially from various sources (e.g. ECMWF and the National Climatic Data Center) in order to provide high quality reanalysis fields. JRA55 is an improved version of JRA25 where many defects are recovered and a more sophisticated forecast system is used. The NCEP reanalysis product which is available from 1979 to present is a continually updating gridded data set (~1.6°×~1.6° with 28 levels), incorporating the numerical weather prediction model outputs with observations using data assimilation techniques. Compared to its previous version, it is a new version that fixes errors and updates parameterizations of physical processes (e.g. surface energy) with more observations being added.

Full comparisons between the reanalysis products are difficult as the quality of the reanalysis dataset is highly dependent on a mix of observations and model forecasts, and also has strong dependence on regions. Comparisons carried out by Lindsay et al. (2014) have shown that the monthly averaged surface variables from ERA-interim including surface temperature, radiative flux, precipitation and wind speed is of greater consistence with independent
observations in the Arctic regions, comparing to that of NCEP and JRA25. Brodeau et al. (2010) has shown that the zonally-averaged shortwave radiation of ERA40 is significantly underestimated between 10°W and 10°E. A more comprehensive comparison is conducted by Chaudhuri et al. (2012) where eight different fields from NCEP, EAR-interim and JRA25 are investigated. In their study it is suggested that no single product is found to agree better in all fields with satellite-derived observations. Precipitation and wind stress fields show significant time-mean and time-variable errors whereas downwelling radiation, air temperature, and humidity fields show small time-mean errors but large time-variable errors, particularly in the tropics.

A brief inter-comparison of seven variables (north and east components of surface wind speed at 10m, air pressure at mean sea level, relative humidity, total cloud cover, precipitation and air temperature) are performed in order to examine the difference between the products. One who might be interested can refer to the references listed in Table 4.6. The comparison is carried out for the year 1998 in which extensive observational data over water columns are available to validate the model results. Fig. 4.20 shows the weekly variation of each field averaged over the model domain. In general all five reanalysis products are mostly comparable to each other due to their similar physical assumptions and assimilation processes, but differences along the timeline are still detected.

The wind speed of NCEP is consistently stronger while the others are of similar magnitude, especially for the v component being showing a typical value of 2-3ms⁻¹ over the year (see Fig. 4.20a and b). This indicates that the NCEP product may produce higher momentum to the ocean, hence more intense mixing. The air temperature shows only small difference in summer whilst a larger difference exists in winter, with a higher value of ~0.8° from NCEP and JRA25. Large difference in the humidity field among these reanalysis products can be
seen in Fig. 4.20e. JRA55 has the highest humidity through the year while ERA-interim is continuously lower than the others. Again larger difference exists in winter than in summer. The total cloud cover, which is a predominant parameter to determine the incoming radiation, also demonstrates significant discrepancy as shown in Fig. 4.20f. This is because that the cloud cover has been found to be one of the weakest features of weather forecasting models compared to other variables (Taylor, 2000). ERA40 and ERA-interim have much denser cloudiness than the others, giving a typical difference of 10%–20% in percentage terms between ERA40 and JRA25 for example. As has been revealed by Brodeau et al. (2010), the zonally-averaged shortwave radiation of ERA40 is significantly underestimated between 10°W and 10°E. It may predict lower downwelling shortwave radiation when using the ERA products. Precipitation is comparable in summer but larger differences in winter can be seen, with greater rainfall from NCEP and JRA products.
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(a) Weekly 10m U-wind (ms$^{-1}$)

(b) Weekly 10m V-wind (ms$^{-1}$)

(c) Weekly mean sea level pressure (Pa)
Fig. 4.20 Weekly variation of seven meteorological fields averaged over model domain for five reanalysis products in 1998. (a) $u$ component of 10m wind; (b) $v$ component of 10m wind; (c) mean sea level pressure; (d) air temperature at 2m; (e) relative humidity; (f) total cloud cover; (g) precipitation
4.3.3 Variations in SST

The statistical errors of SST between model results and remotely sensed data (4km resolution AVHRR Pathfinder Version 5.2) are shown in Fig 4.21. Fig. 4.21a gives the time line of the mean errors in weekly averaged SST from the models driven by five reanalysis products. All models have tendency to underestimate the SST in the majority of time of the year, especially for the ERA products. A similar result to that in section 4.2 is seen that greater differences between models occur in the summer time, resulting primarily from the difference of the total cloud cover (see Fig. 4.20f) as discussed previously. The differences in the mean errors in summer vary between 0.1-1.0°C while a typical of ~0.5°C is found for the rest of time. The SST modelled by the JRA and NCEP data is overestimated (0.3-0.8°C) in summer while it is underestimated by the ERA products as also seen in a ~7km POLCOMS model by Holt et al. (2005) where the same forcing data are used. It is evident that the SST predicted by five models is underestimated by ~0.7 to ~1.0°C from Jan to March, primarily resulting from the inaccurate representation of the initial field. The effects of the initial field for a regional model can last as long as 15 months after which local atmospheric forcing becomes dominant as

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Available online: http://podaac.jpl.nasa.gov/AVHRR-Pathfinder
Fig. 4.21 (a): The time line of weekly averaged sea surface temperature for the year 1998 from models driven by five reanalysis products and remotely sensed data from 4km AVHRR. (b): Root mean square errors of SST between models and AVHRR. (c): Correlation coefficients of SST between models and AVHRR.
revealed by Wakelin et al. (2009). Peak errors are observed in May, reaching -1.25 to -1.75°C.

The RMSe of SST between model results and AVHRR is shown in Fig. 4.21b whilst its correlation coefficient is given in Fig. 4.21c, showing stronger dissimilarity after March when the net solar radiation becomes positive. The model errors exhibit strong time dependence, with the RMSe that can be doubled from one week to another. The RMSe calculated here is sufficiently small, with the maximum value (~0.8°C) even less than the typical error (~1.0°C) reported by Holt et al. (2005). In summer the JRA55 and ERA-interim, with lower RMSe and larger correlation coefficient, demonstrate higher capabilities to predict the spatial distribution of SST than JRA25 and ERA40. This is likely due to improved physics in models and better assimilation techniques used by JRA55 and ERA-interim which are upgraded versions of JRA25 and ERA40 respectively.

The heat budget is also investigated and its viability is shown in Fig. 4.22. The difference in total inward heat flux (see Fig. 4.22b) increases from late May to September, showing a maximum of ~80wm⁻² between ERA40 and NCEP. This is mainly because of the discrepancy in the total cloud cover as indicated in Fig. 4.20. This results in the outcome that the SST predicted by ERA40 is underestimated by ~0.9°C (see Fig. 4.21a). For some periods with extreme cooling, great difference in the heat budget can be also seen, in early April for example. The net heat budget difference in this week is as large as -100wm⁻² between JRA25 and ERA-interim (see Fig. 4.22c). It is, however, not necessarily true that it leads to a large difference in the SST (see Fig. 4.21a) since at the surface boundary of the ocean the heat transformation is a nonlinear process or parameterisation (Elliott and Clarke 1991).
Fig. 4.22 (a): The time line of weekly averaged outward heat flux for the year 1998 from models driven by five reanalysis products. (b): The total inward heat flux (c): The net heat flux.
4.3.4 Variations in water column structure

The vertical structures simulated by the models forced by five reanalysis products are validated against observational data. The observational dataset (Cor98) used here are identical with the study of full model validation described in section 4.1.2, with 13 transects located in the model domain as shown in Fig. 4.23. An example (transect 182) of comparisons is shown in Fig. 4.24. It crosses the whole model domain, capturing the main and complicated features in the sea.

Fig. 4.24a gives the temperature pattern from Cor98 while the others are from model outputs forced by reanalysis parameters. A notable feature which can be observed that the depth of the thermocline from the five models is quite different. Visually, NCEP (Fig. 4.24f) shows better agreement with the observation than the others, giving a more realistic thermocline depth and the bottom fronts. NCEP predicts reasonably the thermocline depth (~30m) compared with the observed, whereas it is found to be much shallower (by ~12m) when modelled by JRA25 and ERA40. It is most likely due to weaker surface mixing associated with lower wind speed (see Fig. 4.20b) as the bottom tidal mixing is identical for all simulations.

At distances from 50 to 100km, JRA25 predicts significantly unrealistic mixing (too weak), leading to much colder bottom water, hence warmer water above the thermocline. A bottom frontal interface at a distance of ~90km is also missed by JRA25 as shown in Fig. 4.24d. NCEP predicts this feature better than the others although the frontal interface is slightly shifted to the left compared with Cor98. The position of the main bottom front is not sensitive to surface forcing, located at ~220km for all model simulations. The sharpness of the secondary frontal boundary for NCEP (~180km in Fig. 4.24f) is more intense than that of
ERA40 and ERA-interim as they tend to be more diffusive. The bottom dome-like water centred from 100 to 150km is less affected, showing a typical temperature of less than 11°C.

**Fig. 4.23** Schematic diagram showing the locations of Scanfish transects in the Celtic Sea.
Fig. 4. 24 The temperature structure of transect 182 from Cor98 Scanfish data and models driven by five reanalysis products.
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Fig. 4.25 Taylor diagram showing the statistical results of temperature between model outputs and Cor98 for 13 transects in Fig. 4.23. Mean errors are included in the brackets with model names.

A more quantitative comparison of the water column temperature is performed by using the Taylor Diagram (Taylor, 2001). It summarises three statistics in a 2D graph, including the standard deviation, correlation coefficient and the RMSe that indicate how closely the test fields match the reference field. The errors between models and Cor98 are calculated from the 13 transects (see Fig. 4.23) and shown in Fig. 4.25. The standard deviation is proportional to the radial distance scaled by the grey arc. The centred RMSe is determined by the distance between the reference point and the model fields scaled by the green arc in the same unit and interval as the standard deviation. The correlation coefficient between two fields is given by
the azimuth indicated by the blue dotted lines. Mean errors are also included in the brackets with the model names.

As seen from the Taylor Diagram, all models tend to underestimate the water column temperature, showing negative mean errors. NCEP exhibits the highest capability in predicting the water column structure with the lowest mean error (-0.39°C), RMSe (1.24°C) and the largest correlation coefficient (R=0.91). However, the mean errors of JRA25 and ERA40 are almost three-fold in comparison to that of NCEP, giving the poorest predictability with larger RMSe. JRA55 and ERA-interim have better predictive capability with reduced mean errors and RMSe comparing to JRA25 and ERA40, but still not as accurate as NCEP.

4.3.5 Variations in stratification

The sensitivity of stratification, characterised by the potential energy anomaly (PEA), the depth of thermocline, the contrast at the interface of the thermocline and the surface-to-bottom temperature difference, are investigated over the model domain. The PEA \( \varphi \) (Simpson and Bowers, 1981) defined by equation (2.1) is used to measure the stratification and mark the frontal positions. It is a vertically integrated measure of energy required to completely mix the water column.

Fig. 4.26 shows the monthly potential energy anomaly in September 1998 for the models forced by five reanalysis products. The patterns of the frontal positions marked by PEA=0 are nearly identical for all model simulations, showing good agreement with previous studies in the Celtic Sea (e.g. Holt and Proctor, 2008; Holt and Umlauf, 2008). However, the modelled strength of stratification exhibits great differences spatially. In the stratified region the strength of the stratification predicted by NCEP, JRA55 and ERA-interim are much stronger than that of JRA25 and ERA40, especially in the Celtic Deep. For instance, the value of PEA from NCEP is higher by 40Jm\(^{-2}\) than that of JRA25 in such region. This is most likely due to
stronger surface heat flux from NCEP, JRA55 and ERA-interim which increases the potential energy, and also due to deeper thermocline depth which maintains higher-level heat in the surface mixed layer. The study of errors quantification of a high-resolution coupled hydrodynamic ecosystem model for the European shelf seas (Holt et al., 2005) demonstrated that the largest errors of the summer time stratification exist in the strong stratified region rather than the frontal region and coastal waters. This, somewhat, gives an indication that the strength of stratification is more sensitive in strong stratified waters.
Fig. 4. Monthly potential energy anomaly in Sep 1998 for models forced by five reanalysis products.
**Fig. 4.27** (a): Surface-to-bottom temperature difference (°C) averaged over the transect along 51.4°N. (b): Depth of thermocline (m). (c): Temperature gradient (°C) at thermocline.
The surface-to-bottom temperature difference averaged along 51.4°N is shown in Fig. 4.27a, reflecting the variability of the stratification along the timeline. In general, the difference starts to increase in May and disappears in December, with the maximum difference occurred in September. This tendency is in a good agreement with the study by Simpson and Bower (1984) where the measurements were conducted at a fixed location in the Celtic Sea. The onset of the stratification ($\Delta T=1^\circ C$) is sooner or later by 4-5 days for different models while the breakdown is more sensitive: two weeks earlier from NCEP than from ERA40. This has been highlighted by Young and Holt (2007) where the timing of the onset and breakdown of seasonal stratification and of its peak demonstrated strong inter-annual variability based on a 40 years simulation in the Irish Sea. Stronger difference exists between June and September, showing a maximum value of $\sim1^\circ C$. Temperature difference modelled by JRA25 is larger than the others from June to August, subsequently drop to the lowest value in September (see Fig. 4.26) when ERA-interim becomes most intensively stratified.

The depth of the thermocline averaged along 51.4°N is compared for five simulations shown in Fig. 4.27b. It is calculated at the maximum of vertical temperature gradient along with the limit $\Delta T>0.5^\circ C$ as used by Holt and Umlauf (2008). It can be seen that the onset of the formation of the thermocline is less sensitive to surface forcing, with the differences in the depth of the thermocline varying from 2-9m. In contrast, the difference reaches $\sim40m$ between NCEP and JRA25 during the breakdown period of the thermocline (see Fig. 4.27b). The thermocline penetrates downwards slowly with time, but with different rates for the five models. The depth of the thermocline from NCEP is greater constantly than those of others by 8 - 10m from August to October, following with marked increases in November.

The temperature gradient at the thermocline, another concern of this study, is also evaluated as shown in Fig. 4.27c. The difference among five models, substantially following the pattern
of the surface to bottom temperature difference (see Fig. 4.27b), starts to increase in June and then reaches the peak point in early September. For example, the temperature contrast at the thermocline for NCEP is ~1°C stronger than that of ERA40 in September, which may result in significant effects on vertical mixing and diffusion at the interface of thermocline. It is then reduced to a similar level in late December.

4.3.6 Discussion and conclusion

The sensitivity of modelled SST, water column structure and seasonal stratification to surface forcing taken from five popular reanalysis products were carried out, showing high correlations of the thermodynamics to the selection of atmospheric forcing. To some extent, the inter-comparison among different surface datasets exhibits dissimilarity along the timeline, especially for the wind speed and the total cloud cover. This results in significant discrepancies in predicted vertical temperature structures and stratification.

The SST predicted by POLCOMS has been validated intensively by a number of studies (e.g. Holt and James 2001; Holt et al., 2005; O’Neill et al., 2012), as well as this model. For this regional model, five reanalysis datasets demonstrate sensitivity in the predicted SST, showing a typical difference in average of ~0.5°C (see Fig. 4.21a) compared with each other. The mean errors against observations were found to be from -0.8°C to 0.8°C in summer, depending on the surface forcing used, while they are 0.5-1.3°C predicted by a ~7km model from Holt et al. (2005) and higher values (0.96-2.6°C) from a ~1.8km model reported by O’Neill et al. (2012). The RMSe of this study (0.2-0.8°C) also demonstrates better skills than the two studies above, the values of which were found to be 0.7-1.8°C and 0.95-1.34°C respectively. This gives the confidence to use any reanalysis product listed in Table 4.6 for the simulation of the mean SST by using this model. However, the results demonstrate strong dissimilarities in the modelled spatial distribution of the stratification indicated by the PEA.
maps (see Fig. 4.26). This gives evidence that great differences in the spatial patterns of the SST exist. It may be attributable to the difference in wind speed, which plays an important role in modulating the spatial pattern of SST through horizontal advection and vertical mixing. So, the spatial patterns of modelled SST might be very sensitive to surface forcing and should be treated carefully.

Compared with the SST, the vertical thermal structure is a primary concern due to its direct impact on sound propagation, placing an extra attention in the model accuracy. The main features of the density distribution in a tidally active sea can be predicted as long as the tidal forcing is provided accurately as examined by Xing and Davies (2001). They also concluded that better representations of the thermal structures, especially in the upper layer, require high-frequency wind forcing as the depth of thermocline can be deepened up to 8m in a daily scale due to its rapid responds to winds. The comparison of the vertical transect in this study also indicates that the modelled vertical thermal structure is highly dependent on the source of surface forcing. NCEP shows the greatest predictability while the others tend to predict shallower thermocline owing to the lower wind speed, hence the weaker momentum. The statistical mean errors in temperature over water columns can be reduced by ~0.85°C when using the NCEP data. ERA-interim and JRA55 shows clearly improvements in comparison with their previous versions (ERA40 and JRA25) based on the statistical results (see Fig. 4.25), however, they are still not as accurate as NCEP.

Looking at the specific features, the difference in the depth of the thermocline is as large as ~12m among different models (see Fig. 4.24), which demonstrates a high sensitivity to surface forcing, owing primarily to the difference in the wind field as seen in Fig. 4.20. The boundary of the bottom front is also a concern since it can change the horizontal propagation angle of sound waves determined by the sharpness of the interface (Weinberg and Clark,
1980). It is shown clearly in Fig. 4.24 that the boundary of the bottom front at ~80km is completely missed by the JRA25 product. This frontal structure is, therefore, changed substantially by using different surface forcing. Furthermore, this is the location where strong density driven currents (>12cms$^{-1}$) exist associated with the large density gradient (see Brown et al., 2004). It is difficult here to explain the misfit of such predicted bottom frontal structure by simply investigating the surface mixing by winds or bottom mixing by tides as the physical processes inside such subsurface frontal system are very complicated. The convection induced by vertical shearing of horizontal density gradients has been found to be very significant to the vertical mixing compared with the shear driving mixing (Holt and Umlauf, 2008). The numerical diffusion, likewise, has the potential to create horizontal diffusion to degrade the sharpness of the interface (see Fig. 4. 24). Accurate representation of such subsurface frontal system requires accurate modelling of mixing, hence the details of the turbulence closure scheme as discussed by Holt and Umlauf (2008) and Holt and Proctor (2008). Although NCEP gives the best prediction in this case, the diffusion (~190km in Fig. 4. 24) is still stronger than the observed. Investigation of the detail of the turbulence model is beyond the scope of this study, but the errors in sound propagation can be quantified by comparing the sound propagation fields calculated from modelled and observed temperature and salinity respectively, which will be presented in the next chapter.

In terms of the stratification, it also shows strong differences among the five models. The strength of the stratification can be as large as 40Jm$^{-3}$ from one model to another in summer as measured by PEA (see Fig. 4.26). Given that the typical error in the modelled PEA by POLCOMS was found to be less than ~6Jm$^{-3}$ in the European shelf seas (Holt et al., 2005), the conclusion is that the stratification is, to a large extent, sensitive to the source of surface forcing. This mainly causes from the discrepancy in the wind field which is a significant contributor to the total mixing in the Celtic Sea (Simpson et al., 1978). For a given volume
averaged heat content in surface layer, the PEA decreases with the decreasing thermocline depth (Simpson and Bowers, 1981), which also explains the shallowest thermocline depth predicted by JRA25 seen in Fig. 4.24. Another potential reason is that the overestimated wind speed by NCEP might introduce stronger mixing in the surface mixed layer, allowing more heat flux to penetrate to the seawater. This mechanism can potentially compensate the degree of the stratification. Compared with the onset of the stratification, the breakdown process is more sensitive to the surface forcing. The results in this study show that the time point for the breakdown of the thermocline modelled by NCEP is two weeks earlier than that of ERA40 (see Fig. 4.27a). As revealed by Luyten et al. (2003), it is often the case that the breakdown of stratification is less well modelled than the onset due to inaccurate prediction in mixing processes. Furthermore, the accuracy in modelling the breakdown of stratification requires more accurate surface forcing since such process is highly dependent on the occurrence of individual storm events.

Different forcing can lead to a typical difference in the depth of the thermocline of ~12m, reaching a maximum of ~40m during the dissolution of breakdown of stratification (Fig. 4.27b). This may cause significant effects on the modelling of underwater acoustic propagation as it changes the main depth of refraction. The surface mixed layer always allows the sound to be ensonified through the water column above the thermocline and reach the surface, leading to more boundary interactions where sound energy losses occur (Lam et al., 2009). The difference of the temperature contrast at the thermocline is also evident particularly in summer, reaching ~1°C in September. This may enhance the inhibition of vertical mixing and diffusion, causing effects on biological processes. Stronger gradient leads to greater Gradient Richardson number, hence reducing the mixing through the thermocline (Pingree, 1980). As shown in Fig. 4.27c, NCEP has the highest potential to limit the diffusion of nutrition out of the bottom water to replenish the nutrients in the euphotic zone. This can
lead to a larger reduction in the primary production in summer since changes in oceanic nutrients are a first order factor in determining changes in the primary production of the northwest European continental shelf (Holt et al., 2012).

Resolution is an important consideration in the choice of surface forcing. Previous studies (e.g. Brossier et al., 2011; O’Neill et al. 2012) have demonstrated that increases in the spatial resolution of forcing can promote the accuracy of models. Note that the surface forcing data used by their studies are produced by the same model respectively. However, this study, which uses surface forcing extracted from different sources, does not show improvements when using higher resolution products. Conversely, NCEP which has the coarsest resolution gives the best prediction. It suggests again that the resolution cannot be considered independently in the selection of meteorological forcing. The inherent quality of the dataset and the parameterisation methods used for the air-sea exchange are of most importance.

The quality of the reanalysis datasets differs significantly from product to product. In this study, the effects of five different reanalysis data on ocean model simulations have been investigated where the model settings are the same apart from the surface forcing, which means that the discrepancies of the model outputs are purely resulted from surface forcing. The NCEP dataset gives the best results, it cannot be, however, concluded that NCEP has the highest capability to drive the water column structures as well as the SST. Other factors can also affect the accuracy of the model. For the model itself, the bulk formulae algorithm which is particularly important for the surface flux parametrisation can introduce errors to the model, which might compensate the errors generated from meteorological forcing. The performance of the reanalysis data is also region-dependent. In regions where more observational data are incorporated the meteorological forcing has higher confidence to drive the variability of the oceanic processes, e.g. in Arctic regions (Lindsay et al., 2014). In this
study, in overall, the NCEP data forced using a version of bulk formulae give the highest confidence to predict the water column structure in the Celtic Sea.

In summary, all reanalysis products have skills to predict the mean SST. However, the predicted vertical structure demonstrates high sensitivity to atmospheric forcing. The difference in the depth of the thermocline is as large as ~12m, to a ~40m during the period of breakdown of the stratification. Strong discrepancy in the modelled mixing and stratification was also revealed. Larger difference in stratification was found in the stratified region, with a maximum value of ~40Jm$^{-3}$ between JRA25 and NCEP measured by the potential energy anomaly. For this work, the NCEP data have been adopted to force the ocean model as they provide the best capability to predict the water column structures.
Chapter 5

5 Coupled ocean-acoustic modelling

5.1 Introduction

There has been a growing interest in the study of shallow water acoustics as the continental shelves are of great economic, biological, social and military importance. The patterns and parameters of acoustic propagation, as well as the level of TL in shallow waters are highly sensitive to the inhomogeneities of temperature and salinity (Katsnelson et al., 2012). The ocean features typical for the shelf sea, such as fronts, eddies, filaments, and variations in the seasonal thermocline, form a highly dynamic environment for underwater sound propagation. Such features in the Celtic Sea were predicted by the ocean model as described in the previous chapters.

In this chapter the high resolution coupled ocean-acoustic model with realistic bathymetry, sediment distribution and meteorological data is used to produce a time varying 3D picture of the temperature and salinity in the Eastern Celtic Sea. The environmental data are conditioned into a range-dependent acoustic model to assess sound propagation and its variability due to a variety of factors, such as the strength of the thermocline, location and strength of ocean fronts, frequency and depth of the transmitter and season of the year, explaining following issues:

i) Error quantification

ii) Seasonal variability of TL
iii) Intra-seasonal (monthly) variability of TL
iv) Intra-seasonal (within one month) variability of TL
v) Hourly variability of TL
vi) Radial transects simulations

**Fig. 5.1** Study area showing the bathymetry of model domain; Transects A, B, C and E are used to calculate 2D transmission losses while D is the location of source for multi-transect calculation; Dots express the locations of sources.

The sound propagation in the Celtic Sea in summer and winter is investigated, demonstrating strong dependence on oceanographic conditions, geographic location of sound source and its depth. In the presence of strong thermocline and the associated subsurface fronts, when the source of sound is on the inshore side of the bottom front the sound energy is mostly
concentrated below the thermocline, showing significant vertical changes in the sound level. When the source is located in the deep stratified water, the vertical changes of sound intensity are weak, but the horizontal propagation distance is strongly shortened due to bottom losses. In contrast, the sound propagation shows little dependence on the location of the source and the sound energy is always evenly distributed over the water column.

The model area covers the sea between 50.08°N to 51.83°N and 7.90°W to 4.00°W (see Fig. 5.1). The details of model descriptions and set-up can be found in chapter 3. POLCOMS is run with full meteorological and tidal forcing for the entire year 2010 to provide hourly temperature and salinity fields for the acoustic model HARCAM. TL calculations are performed over vertical two-dimensional transects, with the vertical grids separated by 1m and 20m in the horizontal to ensure fine resolution. Subsequently, the model is extended to be able to perform multi-transect calculations centred at a point. By doing so, a N×2D model is developed with the capability to simulate 3D TL fields, where N is the number of transects.

### 5.2 Calculation of transmission loss (TL)

TL is the accumulated decrease in acoustic intensity as underwater sound propagates outwards from a source. The common approach to TL calculation is to consider a two-dimensional problem (range and depth), assuming that the azimuthal dependence is small and the source is treated as an omnidirectional monochromatic point (Katsnelson, et al., 2012). In this chapter four 2D transects are used, indicated as A, B, C and E in Fig. 5.1, where the black dots represent the locations of the sources.

Four different transects are selected to represent three different oceanographic and acoustic settings: (i) transect A represents the down slope case condition with the source located in shallow mixed water and the sound propagates into the deeper waters; (ii) transect B
represents the upslope condition with the source in the deeper stratified water. Both transects are utilised to study the seasonal variability of TL in the Celtic Sea; (iii) Transect C is located in a relatively deep part of the sea and used to investigate the intra-seasonal (within one month) variations in TL as the movement of the thermocline and associated front can be clearly seen during the deterioration period of thermocline; (iv) transect E is selected to investigate the intra-seasonal (monthly) and hourly variations in TL since the movement of subsurface frontal system occurs within this transect during the stratification period.

Two omnidirectional sources of sound are considered: the shallow one (at 7m depth) represents a typical large cargo ship and a deep one (at 65, 80 and 85m) is an idealised representation of noise from marine pile-driving activity. Given that ships are low frequency noise sources (10-1000Hz) with the majority of energy concentrated below 300Hz (McKenna et al., 2012) while the peak of the sound level from pile driving activities is often found close to a higher level of 1000Hz (Richardson et al., 1995), the frequencies selected for the TL calculations are, thus, 300Hz and 1000Hz for shipping noise and pile-driving respectively. Note that the locations of the sources are selected close to the main shipping lane in the Celtic Sea identified using the ShipAIS (Shipping Automatic Identification System) data, see BMT, 2013.

Even if the source of sound is omnidirectional, the pattern of sound propagation in a stratified, anisotropic medium such as a shelf sea, can be anisotropic, i.e. depend on the direction of propagation (Hamson, 1997). In order to study the directionality of the sound field, TL is calculated along multiple transects from a single source, see point D in Fig. 5.1. The azimuthal angle of separation between neighbouring radial transects is chosen to be 2.5°. The length of each transect is taken to be a typical 40km in order to fully cover the spatial variations of the oceanic features. All calculations in this study are performed with range-
dependent environmental conditions (2km resolution) obtained from the oceanographic model and real distribution of sediments.

5.3 Results

The calculations in this study show that the main cause of variation in the sound speed is the variation in the sea water temperature. Variations of salinity in the Celtic Sea are small due to lack of significant river discharge. The small impact of salinity, although included in the calculations, is not discussed here.

5.3.1 Error quantification

Before delving into the investigation of TL variability, it is worth estimating how much error from the ocean model outputs is transferred to the acoustic calculations. In order to examine this question, the observational temperature and salinity data and model outputs over the same transects are provided to the acoustic model to perform the TL calculations. The observational data of the Celtic Sea are taken from the Cor98 dataset as introduced in section 4.1.2.3. The transect 189 (see Fig. 4.6) is selected to carry out the comparison, providing a typical stratified water column structure with the subsurface frontal system in the Sea. This error quantification is necessary because the purpose of this study is to investigate the changes of sound level resulting from the variability of the physical environment. The magnitude of errors from ocean model outputs should not affect the conclusions related to the spatio-temporal variations of the environment. Given that the errors of TL are dependent on the frequency and source depth, the TL fields are calculated as a function of frequency at a defined source depth of 65m and a function of source depth at a defined frequency of 500Hz (see the detail in Fig. 5.3).
A 78km propagation distance has been cut from the transect which fully covers the subsurface frontal system. The corresponding sound speed patterns are illustrated in Fig. 5.2a and Fig. 5.2c calculated by equation 3.3, dominated greatly by the temperature structure. The main difference occurs in the margin of the bottom front (~20km), showing more diffuse to the right by the model. Two examples of the TL fields (Fig. 5.2c and 5.2d) are selected to interpret the errors produced by the different input water column data. The red dots in the figure represent the locations of the sources. The TL patterns from Cor98 and POLCOMS are particularly similar visually as seen in the figures. Slight difference can be still seen at the tilted frontal boundary (~35km - ~45km) shown in Fig. 5.2c and Fig. 5.2d. The boundary of the front and the thermocline acting as a reflector separate the sound energy, leading to a step change of sound intensity at the interface (~5dB~10dB).

**Fig. 5.2** (a) and (b): the sound speed of observed and modelled for transect 189. (c) and (d): the corresponding transmission loss at 500Hz. Red dots express the locations of the sources.
A single error evaluation of the TL field may not be representative of overall uncertainties (Kessel, 1999). The TL fields from variable source depths at a constant frequency of 500Hz and different frequencies at a constant SD of 65m are calculated respectively shown in Fig. 5.3. It is obvious that deploying the source in the surface mixed layer demonstrates higher sensitivity to that below the thermocline, reaching a maximum error of ~4dB at SD ~17m shown by the black line. Larger errors are only confined for the source depths at ~17m and 35m, which do not affect the interesting point of this study (7m) in the surface layer, while for the deep sources the errors is negligible. In the case of frequency, the errors exhibit a symmetric pattern, centred approximately at 550Hz. The largest errors occur at 10 and 1150Hz, with a mean value of ~2dB. The errors show an increasing tendency with the frequency indicated by the standard deviation. This is because that higher frequency rays with lower wave lengths are more sensitive to smaller scales of environmental variability in the horizontal direction, causing more uncertainties to accumulate rapidly for the low angle rays.
The errors of TL caused by the input water column data have been investigated for such coupled model above. The TL is more sensitive when positioning the source in the surface mixed layer, especially within the thermocline, which agrees with the experiments conducted by Song et al. (2010). The statistical comparisons indicate that the errors are small (-2dB to 2dB) even in the presence of strong stratification and fronts, except for a few specific conditions. Based on the statistic results obtained, it suggests that such coupled model is capable to predict the TL field over various source depths and frequencies.

5.3.2 Seasonal variation of TL

Transect A

Transect A is located in the southern Celtic Sea near Land’s End (Fig. 5.1). It has been chosen to demonstrate seasonal variations of TL in the setting when the source is located inshore of the subsurface temperature front which develops in summer. The depth of sea along transect A varies from ~30m near Land’s End to ~80m in the north. Calculation have been carried out for two frequencies (300Hz and 1000Hz) with the source deployed at 7m and 20m respectively, which accordingly represents a typical depth of acoustic noise source from a large ship near the surface and pile driving close to the seabed.

The snapshots of temperature, sound speed, and TL are shown in Fig. 5.4 for summer and winter conditions; the snapshots represent daily average conditions on 5 August and 5 December 2010 respectively. Note that the range of temperature and sound speed variations in winter is significantly smaller than in summer.

In summer, a strong thermocline is seen in the deeper areas of the sea, while a subsurface front separates partially mixed coastal waters from the stratified interior, see Fig. 5.4a. The horizontal temperature contrast across the front below 35m is as high as ~3°C. On the onshore side of the front the difference in temperature in the vertical, between the surface and
the 40m level is only ~2.3°C, while on the offshore, stratified side of the front it is as high as ~7°C. In winter, the water column is well mixed vertically with only mild variations of temperature (<0.2°C) in the horizontal direction, see Fig. 5.4b.

The sound speed distribution for transect A is shown in Fig. 5.4c and Fig. 5.4d. In summer, the sound speed pattern closely resembles the temperature structure, due to small contribution from salinity as discussed previously. The horizontal contrast of sound speed across the front is ~9ms\(^{-1}\), whereas the vertical contrast across the thermocline is as large as ~20ms\(^{-1}\). In winter, the sound speed variations are small and do not exceed ~1.5ms\(^{-1}\), mainly due to pressure variations, see Fig. 5.4b.

Fig. 5.4e shows the distribution of sound energy, in terms of TL, for the summer condition, for the source deployed at 7 m depth (shown as a red dot) and generating sound at 300Hz. The acoustic energy is nearly uniform in the vertical within the region between the source and the temperature front (7-9km from the source), and decays rapidly in the horizontal. The pattern changes dramatically on the offshore side of the front, where the acoustic energy is trapped below the thermocline. Beyond the front, the values of acoustic energy above and below the thermocline differ by 15-20dB. This is due to the formation of the near-bottom acoustic duct below the thermocline and deviation of acoustic energy from the surface mixed layer which leads to an enhanced bottom propagation and reduced transmission near the surface. The ocean fronts are not vertical but sloping, and according to Snell's law of refraction this causes the acoustic rays to converge in the near-bottom layer. This pattern is in agreement with the results of Heathershaw et al. (1990) where a ray tracing analysis was used to illustrate the energy disturbance when propagating through an idealised ocean front. Furthermore, the downslope bathymetry can drive the rays into low grazing angles, which not only reduce the bottom loss but also increase the reflection at the bottom of the thermocline.
If the cut-off value of TL is taken as 75dB for example, then the sound (and hence the ship noise) propagates up 40km in the bottom layer but only 13km at the surface.

**Fig. 5.4** Seasonal variations of parameters on transect A: (a) and (b): temperature; (b) and (c): sound speed; (e) and (f) transmission loss at frequency 300Hz with source depth 7m; (g) and (f): transmission at frequency 1000Hz with source depth 20m; Dots represent the locations of the source of sound at a depth of 7m and 20m respectively.
In winter, the TL pattern is much more uniform in the vertical than that of summer, see Fig. 5.4f. The range of sound propagation at the surface is significantly greater than in the summer and reaches 40km (at 75dB cut-off). Near the bottom the range is 31km, i.e. slightly shorter than in the summer. Sound energy slightly converges in the surface layer due to a weak surface acoustic duct seen in Fig. 5.4d.

The effect of variation in the sound source frequency from 300 to 1000Hz and source depth from 7m to 20m on the transmission loss is seen in Figs 5.4 (e-h). The general pattern at 1000Hz is similar to that of 300Hz both in summer and winter with the main difference being a greater transmission loss at 1000Hz. In summer, the sound energy level at 1000Hz in the upper layer in summer is lower by ~15dB compared to 300Hz, whereas it is only ~3dB lower in the bottom layer. The difference between the surface and bottom layers is also greater at the higher frequency reaching ~20dB in the far field in the summer and ~10dB in the winter. In winter the TL at 1000Hz increases (and hence the sound level decreases) along the range more rapidly compared with that of 300Hz, showing ~90dB loss in 40km.

Fig. 5.5 shows the TL vs range at a depth level of 10m for the TL fields presented in Fig. 5.4. The seasonal difference (summer vs winter) increases with distance from the source for both lower (300Hz) and higher (1000Hz) frequencies reaching ~16dB at 35-40km from the source. At 40km range, the minimal loss of sound energy (68dB) is achieved in the winter at 300Hz, and the maximal loss (100dB) occurs in the summer at 1000Hz due to concentration of sound in the near bottom duct, as well as stronger bottom absorption at higher frequencies.
Fig. 5. 5 Transmission loss along transect A with receiver depth at 10m, other parameters are as in Fig. 5.4.

Transect B

The transect B in the north-eastern part of the Celtic Sea is used to analyse the sound fields when the source (shown as a black dot in Fig. 5.1) is located offshore of the temperature front. The source frequencies are 1000Hz and 300Hz, deployed at 80m and 7m respectively.

Similar to transect A, in summer a strong thermocline is developed in the interior of the sea and a subsurface tidal mixing front is formed closer to the coast, leading to a horizontal temperature contrast of ~4°C at a depth of 50m. A cold bottom water pool is seen in summer, with temperature <11°C, see Fig. 5.6a. The temperature pattern is very different in winter, when water is well mixed both vertically and horizontally from surface to the seabed, with very small horizontal variations not exceeding 0.4°C, see Fig. 5.6b.

In summer, a surface acoustic anti-duct (high values of sound speed) is formed, and the variations of sound speed across the transect reach ~15ms\(^{-1}\). In winter, the sound speed
distribution is nearly uniform with a maximum variance of ~2ms$^{-1}$, forming a weak surface duct, see Fig. 5.6d.

The pattern of TL (Fig. 5.6e) from the source located at 7m depth offshore of the front is significantly different from that seen in Fig. 5.4, when the source was onshore of the front. Due to upslope topography, the downward looking acoustic rays resulting from intense refractions at the horizontal thermocline interface meet the bottom at higher grazing angles, which results in greater absorption by sediment. As a result, the acoustic energy is not concentrated in the bottom layer but is distributed vertically nearly evenly. The acoustic range (at 65dB cut off) is 11km at the surface, i.e. similar to Transect A, but only 8km near the bottom, much smaller than in Transect A (where it is ~20km).

In winter (Fig. 5.6f), the sound energy emitted at high grazing angles is absorbed rapidly by the seabed within a short range due to upslope topography, whereas the low angle rays can propagate further away. This pattern is supported by a weak surface acoustic duct, which can trap the rays in the surface layer. Acoustic energy propagates further horizontally before reaching the bottom, reducing the interactions with sediments and thus the bottom losses. This leads to much better propagation in winter, showing a surface range (at 65 dB cut-off) of 25km and a near-bottom range of 16km.

The pattern of TL when the source is located near the bottom, at 80m depth is shown in Fig. 5.6g and 5.6h. This situation is more applicable to the noise produced by marine pile-driving rather than by shipping as in previous examples.
Fig. 5.6 Seasonal variations of parameters on transect B: (a) and (b): temperature; (c) and (d): sound speed; (e) and (f) transmission loss with source depth 7m and frequency 300Hz; (g) and (f): transmission loss with source depth 80m. The frequency of sound is 1000Hz. Dots represent locations of the sources.
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Fig. 5.7 Transmission loss along transect B with receiver depth at 20m, other parameters as in Fig. 5.6.

In summer, the sound energy propagates mainly within the near-bottom acoustic duct formed by cold bottom waters below the thermocline. To the contrary, the sound energy in the upper warm layer is reduced compared to the case when the source is at 7m, leading to a strong vertical difference in the sound level. This effect is only seen at ranges greater than 4-6km, as only the acoustic rays at high grazing angles are capable of piercing through the thermocline. Due to formation of the near-bottom duct, the TL in the lower layer is higher up to ~15dB than near the surface. Acoustic rays at low grazing angles are trapped between the bottom and the thermocline, are subject to a reduced absorption by the sediment due to large incident angle, and therefore propagate over a greater range. The propagation range (at 75dB cut-off) is up to 40 km near the bottom, and only 8-9 km near the surface. The temperature front has an effect on sound propagation, allowing a leakage of sound energy from the bottom into the upper layer beyond 30km range from the source.
In winter, the pattern of TL is different to the case of the shallow deployment of the source (7m). The propagation range is shortened significantly due to increased sound frequency. Since the water column is well mixed, the seasonal (summer vs winter) difference in TL in the bottom layer (below 50m) is up to ~20dB.

Fig. 5.7 shows the TL at a depth of 20m for transect B. In winter, the difference in TL between the shallow (7m) and deep (80m) sources is significant. It increases with range and reaches ~15dB at 40km from the source. In summer, the TL of deep source is ~5dB higher at short range (<20km) while it is converse at the far field with the shallow source predicting a larger TL of ~8dB. The lowest propagation occurs in winter for the low frequency and shallow source (e.g. shipping noise), giving a seasonal difference in TL of ~16dB.

5.3.3 Intra-seasonal (monthly) variability of TL

In order to study the monthly variations of TL resulted from changes of water column structures, transect E (see Fig. 5.1) is selected where the influence of slope is reduced due to relatively flat bathymetry. In addition, it is the location where the seasonal front extends most eastwards and persistently exists during the summer time. The development and retreatment of stratification and the movement of fronts can be observed in transect E. The monthly variability of temperature is shown in Fig. 5.8, in which the first two rows are the mean temperature in degree from May to October while row 3 and 4 are the corresponding standard deviation. Such time period is chosen since it covers the development, enhancement and retreatment of the front and thermocline, capturing the major variations. Note that the scale of the colour bar of the mean temperature is different for clarity in Fig. 5.8.
It can be seen that the thermocline penetrates downwards from a depth of \( \sim 15\text{m} \) in May to \( \sim 30\text{m} \) in September, and disappears in October as shown in Fig. 5.8. Meanwhile, the contrast of the temperature between the surface mixed layer and the bottom layer increases from \( \sim 0.8\degree\text{C} \) in May to the maximum (\( \sim 4.5\degree\text{C} \)) in August, following a slight reduction in September. The movement of the bottom front is evident, extending eastwards from May to August and retreating westwards in September. The corresponding standard deviation marks the strong spatial variation as indicated by the figures. More variability occurs in the surface mixed layer in June while strong variations exist along the boundaries of the bottom cold water in July and August. In contrast, greater changes arise in the cold bottom water in September.

In order to evaluate the spatial variability of the TL of the transect when sound propagates in such stratified water, the Empirical Orthogonal Function (EOF) is utilised as it is a good approach for analysing the spatial and temporal variability of a single field, especially for
long-term variations. In this case, the TL is calculated for every hour which is the output frequency of POLCOMS.

The EOF method decomposes the overall variance into different uncorrelated patterns by solving the eigenvalue and eigenvector problem, and the importance of each pattern is measured quantitatively by the eigenvalue solved. Once the time series data (hourly TL for this case) is ready, the EOF approach starts by forming the covariance matrix $\Sigma$:

$$\Sigma = \frac{1}{N-1}X^TX$$

(5.1)

where $X$ is the data matrix and $N$ is the number of time series. Each two dimensional TL map (Fig. 5.6g for example) is rearranged to a single array and assigned to fill the row vector of $X$ while the column is the time series for each grid point. The eigenvalue and eigenvector of $\Sigma$ can be solved by:

$$\Sigma C = C\lambda$$

(5.2)

where $\lambda$ is a diagonal matrix comprising the eigenvalues of $\Sigma$, and $C$ contains the eigenvectors of $\Sigma$ corresponding to the eigenvalue $\lambda_i$ (hereafter referred as EC). Each eigenvector (hereafter referred as $EOF_i$) obtained can be reconstructed back to the original 2D transect, representing the pattern of spatial variation, the eigenvalue $\lambda_i$ of which gives the importance quantitatively and usually is normalised to 1 or 100. This approach allows to find the largest $\lambda_i$, the corresponding $EOF_i$ of which can be, thus, used to characterise the biggest spatial variations. The projection from the anomaly field $X$ onto the $EOF_i$ can be obtained by the equation:

$$\tilde{a}_i = X \times EOF_i$$

(5.3)
where $\tilde{a}_i$ is known as the principal component (PC) of the $i^{th}$ EOF mode, providing the oscillations in time series. As the EOF is an algebraic approach, the $EOF_i$ is dimensionless, giving the difficulty of interpretations in terms of quantities. Instead, a variance map is utilised here, which is defined as the square of correlation coefficient between $\tilde{a}_i$ and the original TL values at each grid point. The correlation coefficient between two fields is calculated using equation 3.26.

The variance map displays the spatial distribution of variance in percentage explained by the $EOF_i$ mode, ranging from 0 to 1. A lower value means weaker variation while a higher indicates stronger variability. In practice the first few largest $EOF_i$ can always capture the major dynamics of the field, for example accounting for greater than 90% anomaly measured by $\lambda_i$.

**Shallow source 7m:**

Fig. 5.9 presents the variance maps of TL with a frequency of 300Hz and SD of 7m (the case of shallow source) for each month. Two modes (EOF1 and EOF2) are selected since they account for a percentage of at least greater than 80% as indicated by the values of EC in the maps. In May the thermocline and front is shallow and weak shown by the mean temperature in Fig. 5.8. The TL variation in May (Fig. 5.9) is dominated by the first mode, accounting for 94.5%. Also, one can see that the low angle rays can propagate far from the source before encountering the seabed due to the weak thermocline, hence weaker refraction. It exhibits strong spatial variations, almost over the whole transect excepting the propagation distance less than ~8km. In June the thermocline and bottom front is enhanced in which the temperature of the surface layer increases more rapidly than that in the bottom water, as illustrated in Fig. 5.8. The resulted TL anomaly indicates that the spatial variations act similarly to those of May, but the low refraction pattern disappears compared with that in
May. The changes of TL between 0 and ~8km are small whereas it increases along the travel distance after ~8km. An extremely weak surface duct exists in June as identified by the second mode (EOF2), occupying only 4.3%. The surface mixed layer is strongly heated in July and the most intense variations occur in the thermocline and the associated margin of the front. However, the cold bottom pool and the water mass on the right of the front only warm up mildly. Turning to the TL variability in July, the first model (93%) reveals stronger variance in the surface duct and significant variations exist on the right of the cold bottom water. The TL in the cold bottom is, however, less affected.

In August, the thermocline penetrates to the deepest depth when the strongest surface heat flux exists. The front extends most eastwards whilst the water on the east of the bottom front is well mixed. As illustrated by the standard deviation map in Fig. 5.8, the surface mixed layer and bottom mixed layer remain relatively steady, the margin of the thermocline and front, however, varies dramatically. Looking at the TL variations in August, two clear patterns are identified by the variance maps. EOF1 gives the strong surface duct anomaly accounting for 64.1% whilst EOF2 presents 27.2% variations occurring on the right of the bottom front, likely due to the perturbation of frontal position. For the other locations, the changes of TL are particularly small. Due to the decreased heat flux in Sep, the mechanical energy from tides and winds can enhance the mixing, hence shifting the front to the deeper water. Meanwhile the cold bottom water becomes warmer and the temperature in the surface layer is decreased slightly (see Fig. 5.8). As the majority of water is well mixed, the low refraction propagation pattern is evident as shown by EOF1. Strong variations occur almost over almost the whole transect, occupying 88.4%, with more noticeable variations occurring below the thermocline. Compared with August, the surface duct in September is remarkably weakened to only 10.7%. In October variations of TL are small due to well mixed conditions.
**Fig. 5.9** EOF1 and EOF2 for the shallower source 7m with a frequency of 300Hz. EC represents the principal component of each EOF.
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Deep source 60m:

Fig. 5.10 gives the variance maps of TL with deeper SD (60m) with a frequency of 1000Hz. In May the first model occupying 80.2% variability exhibits significant variations over the entire bottom water, whereas the TL changes are more pronounced before the location of front in the surface layer. The second mode is low, accounting only for 10.2%. In June the TL variation in the surface layer increases along the range while in the bottom water a cluster of energy is concentrated after ~8km with interactions between the thermocline and the bottom, then released after the front. The overall spatial variability is remarkably reduced compared with that of shallow source (10m) whereas the second mode (EOF2) displays a stronger pattern (27.7%). In July and August variations are concentrated mainly near the boundaries of the thermocline and surface layers in the well mixed regions. In September strong variability near to the source (<~7km) and in regions after the bottom fronts (>~25km) can be observed, whereas in between weaker variations exists characterised by EOF2 (26.9%). In Oct the degree of variability for both modes is similar to that of the shallow source, but with minor variation occurring in different locations.

It can be seen that the TL in such waveguide displays highly spatial and temporal variability from May to October. The overall variations are summarised as follows:

1. Demonstrate high correlations of source depth and range. The variations increase along with the range as the uncertainty can be accumulated more quickly in the horizontal direction.

2. Since the horizontal thermocline and the vertical subsurface frontal boundary separate the transect into four subsections, each presents different variability.
3. The variation of TL for the SD at 7m is dominated by the first EOF mode while the varying pattern is more complicated for the case of SD at 60m. The second mode still plays a significant role in characterising the spatial variations for the deeper source.

4. Large variations normally occur near the thermocline and the mixed water column located on the right of the bottom front, the bottom cold water however demonstrates weak variability of TL.
Fig. 5.10 EOF1 and EOF2 for the source depth at 60m with a frequency of 1000Hz. EC represents the principal component of each EOF.
5.3.4 Intra-seasonal (within one month) variability of TL

Transect C (Fig. 5.1) is located in the deeper part of the Celtic Sea, in the area subject to rapid changes of the ocean front and it is used to study the intra-seasonal (within 1 month) variability of the transmission loss. For this purpose 2 dates (1 and 26 October) within the autumn period of fast deterioration of the thermocline are chosen. Fig. 5.11 is similar to Fig. 5.4 and Fig. 5.6, with the only difference that the left and right hand panels are separated by only 3.5 weeks (between 1 and 26 October) rather than four months.

As seen from Fig. 11a and Fig. 11b the thermocline and the subsurface front are only partly eroded during the period from 1 to 26 October 2010. However the resulting change in the TL is very significant, especially in the near bottom water. The TL at a receiver depth of 55m is then compared shown in Fig. 5.12. The maximum difference in TL (~10dB) is achieved at a medium range of 15-20km. Beyond the 26km range, the difference is negligible (see Fig. 12). The convergence of the TL curves beyond the 26km mark is due to the effect of the coastal temperature front, which causes a stronger refraction of acoustic energy upwards and entering the beyond-the-front area of vertically mixed waters. The difference disappears completely beyond the 26km range.
Fig. 5.11 Intra-seasonal variations of parameters on transect C: (a) and (b): temperature; (b) and (c): sound speed; (e) and (f): transmission loss with source depth 85m. Sound frequency is 1000Hz. Dots represent the locations of sources.
5.3.5 **Hourly variability of TL**

The seasonal and intra-seasonal effects of water column variations on TL are studied by extracting selectively the snapshots and the EOF method is also adopted to evaluate the monthly TL variations caused by the varying thermocline and frontal system. However, the EOF is a statistically averaged approach. Some periodic effect might be, therefore, excluded using such method. The EOF gives only a relative measure of importance, with no absolute values presented. For such temperature-dominated tidal mixing frontal system, the frontal position is strongly dependent on the tidal propagation but less sensitive to the surface heat flux (Holt and Umlauf, 2008). Accordingly, the periodic tides can introduce perturbations of the frontal position, July and August in Fig. 5.12 for example. As the thermocline splits the water column into surface and bottom layers and the bottom front separates the transect into left cold bottom pool and right well mixed water (see Fig. 5.8) and each of subsection demonstrates strong variability as seen in the case of EOF study. Four locations in transect E are, thus, selected to extract the hourly TL data, namely at the surface of 13km, at the bottom of 13km, at the surface of 35km and at the bottom of 35km. For each location, the TL data are averaged over a rectangular box with vertical 10m and horizontal ±500m. The hourly TL

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**Fig. 5.12** Transmission loss: source depth 85m, receiver depth at 55m for Fig. 7e and Fig. 7f.
data extracted from these four locations are plotted against time for both the shallow and deep sources.

*Source depth at 7m:*

The TL along the timeline for SD at 7m is illustrated in Fig. 5.13a, in which the lighter background lines are plotted with the original hourly TL data and the bold solid lines are the average over 12 hours close to the semi diurnal tidal cycle 12.42 hours. Fig. 5.13b gives the SST averaged over the transect. It can be seen that the TL variation at 13km varies mildly, giving slightly lower losses in the surface due to the refraction bending as identified in Fig. 5.6e. However, in the far field (35km) the TL changes dramatically, following the trend of SST dominated by the surface heat flux. The TL increases gradually with the increasing surface temperature, resulted from the enhanced temperature gradient in vertical. The difference of TL between different seasons can reach as large as ~23dB. An interesting feature can be observed that a decrease in temperature can introduce a large reduction of TL as indicated in Fig. 5.13, particularly at the surface at 35km. The TL at the temperature turning point has been investigated (not shown here) and this can be explained by the fact that a decrease in temperature in the surface can introduce a positive sound gradient vertically and generate a surface acoustic duct, significantly reducing the TL in the upper layer. These results reveal a fact that the heating and cooling processes in the upper layer are different, correspondingly causing different impact on TL. Comparing the original hourly data (lighter colour) to the 12-hour averaged TL, the tidal oscillations can be seen clearly but with relatively lower amplitude (~4dB), which suggests that the TL is more sensitive to the surface heat flux than to the tides for sources deployed in the surface mixed layer.
Fig. 5.13 (a) hourly propagation losses at different four locations of transect E for source depth at 7m; (b) the sea surface temperature averaged over transect E.

Source depth at 60m:

Fig. 5.14a represents the TL scenario of SD at 60m whilst Fig. 5.14b shows the currents velocities averaged over the entire transect. In the bottom layer, TL at both 13km and 35km exhibits weaker variance and mild tidal perturbation. In contrast, in the surface layer at 13km (red) the TL starts increasing from June due to increased heat flux meanwhile the tidal signature becomes more pronounced, giving a TL difference of ~10dB vertically. At the far position of the upper layer (35km), strong semi-diurnal tidal influence is seen, with averaged amplitude ~9dB. Compared with the case of shallow source (7m), the tidal effects are more
significant by placing the source below the thermocline, which is coincident with the study by Xu et al. (2009). When the source is located below the thermocline, the cold bottom water can trap a large amount of low angle rays. The position of front is, thus, critical and sensitive to the trapped energy. It directly affects the redistribution of energy after the bottom sound energy propagates across the front. Note that the spring-neap cycle of tides can lead to the adjustment of frontal positions (Simpson et al., 1980), typically 2-3km.

The averaged velocities over the transect are given by Fig. 5.14b, from which evident spring-neap cycles are presented. The bold black line in Fig. 5.14a, over which the semi-diurnal effects are excluded by averaging, displays clear spring-neap variations following the kinetic energy pattern (Fig. 5.14b) and giving a mean amplitude of TL ~7dB. Variations generated by tides and heat flux decrease from the middle of September as the front and thermocline are weakened.
Fig. 5. 14 hourly propagation losses at different four locations for source depth at 60m; (b) the currents velocities averaged over entire transect E.
5.3.6 Radial transects simulations

The radial variability of sound propagation is calculated for the source located at point D, (see Fig. 5.1) by computing multiple vertical TL transects with azimuthal resolution of 2.5°. The results are shown for two seasons and two depths of the source: 7m (mimicking the ship noise, Fig. 5.15) and 65m (mimicking pile-driving noise, Fig. 5.16).

Fig. 5.15 Radial graph of transmission losses from an omnidirectional source deployed at 7m and receiver at 12m (a,b) and 50m (c,d) depth for the summer (a,c) and winter (b,d) conditions.

Fig. 5.15 shows the transmission losses as a function of season (August vs December) and water depth of receiver (12m vs 50 m). In summer, the sound energy spreads more efficiently (lower values of TL) in the NNE and SSW directions in the upper layer. In the lower layer
there is a clear tendency to propagate to the west into the deeper part of the sea. The directionality of sound propagation is a combined effect of variations in temperature (and to a smaller extent, salinity) of the water column, sediment, bathymetry and sea surface state. In winter, the sound propagates in a virtually axisymmetric pattern and has greater ranges both in the upper and lower layers than in the summer.

The difference in the propagation range between two seasons can be as high as ~25km, depending on the direction, water depth and specific threshold. For instance, transmission loss of 80dB is achieved at 40km range in winter while the same loss occurs at only at ~15km in summer along some directions.

In summer, the horizontal distribution of TL shows significant azimuthal variability both in the upper and lower layers, when the source is located below the seasonal thermocline (at 65 m) – the situation is more typical for the pile-driving activities (see Fig. 16). In contrast to the case of a shallow source deployment (more typical for ship noise) the sound energy tends to be directed towards the coast both in the upper and lower layers, see Fig. 16a and Fig. 16c. The difference in the TL and hence the sound level in different directions can be as high as 30 dB at the same range from the source.
5.4 Discussion and conclusion

The results of the combined ocean-acoustic modelling in the Celtic Sea show that the pattern and range of sound propagation depends strongly on the 3D distribution of temperature and salinity, and hence is highly variable both in time and space. The results show a clear influence of thermocline and subsurface front on sound propagation. In summer, a sharp thermocline and a subsurface temperature front make the propagation of sound energy very different from the winter conditions, when the water column is well mixed. The warm layer of water above the thermocline creates an anti-duct which refracts the sound energy into the
lower layer in agreement with previous studies by (Lindar and Gawarkiewicz, 2006; Lam et al., 2009). The subsurface temperature front and the slope of the seabed contribute to the azimuthal variability of the sound propagation. When the source is located on the inshore side of the front, in vertically mixed waters, the sound energy concentrates in the near-bottom layer even when the source is near the surface, see Fig. 5.4e and Fig. 5.4g. The temperature front typically has a slope between 1:100 and 1:1000, see (Fedorov, book on fronts), and hence causes refraction of acoustic rays with low grazing angles away from the surface layer.

The thermocline separates the water column into a two-layer upper system: warm upper layer (high sound medium) and cold bottom water (low sound speed), see Fig. 5.4 and Fig. 5.6. The effects of thermocline on sound propagation are well known and characterised by the mechanisms of reflection and refraction. This can be explained by applying the Snell’s law. When the source is placed in the warm layer, strong refraction occurs at the interface of the thermocline. This leads to an increased grazing angle of rays at the interface of the seabed, hence stronger bottom absorption. Such effects reduce the propagation distance up to tens of kilometres in summer as shown in Fig. 5.6e. When the source is located below the thermocline, reflection at thermocline is dominant where sound travels from low speed media to a higher one. Sound rays are reflected greatly by the thermocline, resulting in much higher energy level in the bottom cold water in summer (see Fig. 5.6g).

The effect of bottom fronts on sound propagation is more complicated and has been investigated by a number of studies (e.g. Weinberg and Clark, 1980; Heathershaw et al., 1991; Abbot et al., 2001). An experiment performed by Weinberg and Clark (1980) revealed that the change of the refraction angle exceeds 1°C when sound propagates across a front, dependent on the sharpness of the front and the relative distance between the source and the front. This highlights the potential effects of fronts on sound propagation. By altering
artificially the frontal structures when sound propagates through the front, a change in the sound level of ~8dB was found by Abbot et al. (2001). In this study, the changes of the sharpness of the subsurface front and its movement are shown clearly (see Fig. 5.8 and Fig. 5.11). The results from this study show that the sound propagation is very different when crossing different frontal systems. The bottom fronts allow leakages of sound energy trapped in the cold bottom water to redistribute over the water column (Fig. 5.11). Changes in the frontal structure (e.g. position) then cause impacts on the distribution of the sound energy on the right hand side of the front where the water is well mixed (see Fig. 5.11e and f). Greater changes in the sound level always occur in the upper layer as indicated by the black line in Fig. 5.14a, showing spring-neap tidal signals.

The combined effects from thermocline and fronts on sound propagation demonstrate high dependence on the geographic location of the sources. When a shallow source is located on the seaward side of the front, the acoustic rays meet the frontal gradient at higher grazing angles, and are absorbed by the bottom sediment more efficiently. When the source is located near the bottom, inside the acoustic channel located below the thermocline, the grazing angles get smaller and sound propagates over longer ranges near the bottom, see Fig. 5.6g. For a shallow source on the inshore side of the front, the range of sound propagation in the upper layer is greater in winter than in summer; however in the near-bottom layer it is greater in summer, see Fig. 5.4. Conversely, for the same source located on the seaward side of the front, the range both in the surface and bottom layers is greater in winter, see Fig. 5.6. Comparison of TL from the same source on 1 and 26 October 2010 shows that variations of the temperature field within 3.5 weeks are sufficient for significant changes (~10dB) in the sound propagation, see Fig. 5.11.
Previous studies focused generally on the impacts of the water column structures on sound propagation. The indirect causes (e.g. surface forcing), which are important to a coupled ocean-acoustic system, received less attention. The results in this study show that the variability of TL is highly dependent on the physical forcing (e.g. heat flux and tides). The shallow sources (e.g. ships) are more sensitive to the surface heat flux (see Fig. 5.13) as their propagation are very dependent on the strength of the thermocline as discussed previously. In the Celtic Sea, the upper layer responds to surface forcing very rapidly while the bottom water is always less affected. The strength of the thermocline is, therefore, greatly determined by the surface heating, which changes the vertical temperature gradient quickly. Note that the cooling process in the surface layer, which is always more rapid than heating, can result in a sudden reduction in the level of TL (Fig. 5.13). Tides play an important role for deeper source (e.g. pile driving) in determining the TL variability, which agrees with the study by Xu et al. (2009). It is most likely due to the changes of the position of the subsurface front. Fig. 5.14a demonstrates clear spring-neap variability of TL, showing a typical magnitude of ~7dB that is comparable to the study (~8dB) by Abbot et al. (2001). Furthermore, the spring-neap cycle of tides causes an adjustment of tidally frontal position of 2-3km (Simpson et al., 1980). Such adjustment substantially changes the relative distance between the source and the front, which modify the propagation patterns as revealed by Weinberg and Clark (1980).

A large amount of sound energy below the thermocline (see Fig. 5.20c), when the source is close to the bottom, should be taken into account in planning pile driving activities.

Biologically, marine mammals radiate sound as primary functions for communication, echolocation and foraging (Richardson et al., 1995). Low frequency sounds from some animals, for example, the baleen whales, can propagate over hundreds of kilometres (Richardson et al., 1995; Clark, et al., 2009). This study suggests that in summer the
communication range could be reduced by ~27km when the animals stay in the surface mixed layer (Fig. 5.6). In contrast, when marine mammals dive to the bottom layer, their range of communication is enhanced due to the thermocline and fronts. In addition, noise from anthropogenic activities (e.g. pile driving, seismic survey and ships) has received significant attention due to their detrimental effects on marine animals (Southall et al., 2007). The concentration of sound energy in the lower layer in summer from shallow sources (e.g. large ships) may have an adverse effect on marine animals. Near the bottom, the animals could be exposed to shipping noise in a larger area than at the surface.

Estimating the source level (SL) of marine mammal calls is an important biological measurement, which can be calculated normally using the sonar equation: SL = RL + TL, where RL is the received level obtained using hydrophone array measurements. TL is, thus, of particularly importance to determine the SL. The generic spreading models were broadly used in biological areas, which is a time-depth-independent logarithm dispersion. The full model study reveals that the difference in the level of sound energy in time and depth can reach ~10 - ~35dB. When the sound travels through the front, the descending tendency of TL can be broken and switch to an ascending propagation as shown in Fig. 5.12. The geometric spreading models are, thus, hard to capture such complicated TL variations. The time and environmental dependence must be, thus, considered in this application as the correlation scale is large.

The main conclusions are as follows:

- The errors of TL between simulations using observational temperature and salinity and model outputs are quantified, giving the confidence that the ocean model is sufficient enough to provide the input data for the acoustic model (see Fig. 5.4 and Fig. 5.7).
• In summer, when the source of sound is on the inshore side of the front, the sound energy is mostly concentrated in the near-bottom layer (between seabed and 20m level). In winter, the sound from the same source is distributed more evenly in the vertical. The difference between the sound level in summer and winter at 10m depth is as high as 15dB at a distance of 40km from the shallow source in the offshore direction.

• When the source of sound is on the seaward side of the front, the sound level from shallow source is nearly uniform in the vertical and the transmission loss is significantly greater (~16dB at 40km distance) in the summer than in the winter. In contrast, sound energy from deep source is trapped in the bottom cold water, leading to a much lower transmission loss (~20dB) in summer than in winter, or ~10dB fluctuation during the deterioration of the thermocline in late autumn.

• The surface forcing highly affects the TL for shallow sources whilst the tides can introduces significant variations of TL for deep sources. The temporal variations of TL can reach as large as ~20dB.
Chapter 6

6 Seasonal changes in shipping noise exposure experienced by diving seals

6.1 Introduction

Shipping noise is a major contributor to anthropogenic noise in the sea, which is now classed as a pollutant in accordance with the MSFD. However, little is known about its spatio-temporal variability and how it impacts marine organisms. Animals have different hearing characteristics and sensitivities to noise. Seals are considered to be one of the most vulnerable species to shipping noise as the sound frequency band of seals is overlapped by that of ships (Southall et al., 2007). Grey seals (*Halichoerus grypus*) are chosen here to investigate the likely noise exposure due to a number of reasons. The frequency band of the hearing of grey seals is large, ranging from 200Hz to 200,000Hz (Richardson *et al.*, 1995), while the dominant frequency band of shipping noise is between 10-1000Hz. There is clear indication that noise produced by ships can be perceived by grey seals. They are able to travel large distance from one foraging ground to another, which increases the risk of being exposed to a noisy environment. Shipping noise thus has potential to affect grey seals in a number of ways (e.g. masking and behavioural disturbance). Furthermore, grey seals are benthic foragers and they generally dive to the bottom for foraging and return to the sea surface for breathing. As revealed in chapter 5, the noise level is highly dependent on the water depth. This makes seals have high potential to experience step changes of sound during their descent/ascent.
through the water column. In addition, there are large diving datasets available, which offer the possibility to predict the noise level during their diving.

Acoustically, the adverse effects of anthropogenic noise on marine life are assessed normally by relying on the absolute received level (RL). There have been a number of sound exposure metrics to characterise sound intensity. Due to the nature of long-term and non-pulse sound signal of shipping noise, Sound Exposure Level (SEL) is considered as an appropriate exposure metric for marine mammals (Southall et al. 2007) since it considers the chronic effect and is a measure of the accumulative effect of sound energy over a certain duration. RL and SEL are, thus, used in this study to characterise quantitatively the sound energy received by seals.

The generic spreading models are an efficient way to simply assess the TL fields and used widely for noise applications (e.g. Hatch, et al., 2008; Bailey et al., 2008; Erbe et al., 2012). They are defined by $TL(r) = N \log(r) + \alpha r$, where $N$ is a factor for spreading loss, $\alpha$ is the absorption coefficient and $r$ is the range. $N$ is defined by 10, 20 and 15, representing the cylindrical spreading, spherical spreading and intermediate empirical spreading respectively. The absorption coefficient is estimated using formulas from Francois and Garrison (1982). The spreading models may be insufficient to predict noise patterns in shallow seas due to their simplicities. A number of studies (e.g. Hatch, et al., 2008; McKenna et al., 2013) have placed emphasis on the incorporation of the detailed environmental conditions into shipping noise estimation in shelf regions due to their significant impacts on sound propagation.

Despite incorporating complicated environmental factors into shipping noise estimation has been documented as an important element, a few data or studies are only available to evaluate the potential impacts on marine organisms and spatial planning. In this chapter the seasonal changes of noise generated by a cargo ship are studied and the differences of TL predicted by
the full model and that calculated by generic spreading models are compared. The potential shipping noise exposure experienced by grey seals in the Celtic Sea is then investigated by overlaying their GPS tracks and dive data, over a state-of-the-art ocean (POLCOMS) and acoustic (HARCAM) propagation model populated with real-time AIS shipping data in summer and winter.

6.2 Materials and methods

6.2.1 Study area

The Celtic Sea shown as the colour map in Fig. 6.1 is selected as the study area, located in one of four MCZs in UK waters. This area is of great ecological importance to seals, harbour porpoises, common dolphins, bottlenose dolphins and minke whales since abundant species have been observed in the region (Reid et al., 2003; Hammond et al., 2008). Geographically, the Celtic Sea is encompassed by the coasts of southern Ireland, south-west Wales, Cornwall, and by the St George’s Channel, the English Channel and the 200m isobath of the continental shelf linking with the North Atlantic (Thompson, 1986). Traffic in this region is heavy as it links the Atlantic with UK coastal waters. Shipping routes in the Celtic Sea mainly connect outer waters with the Irish Sea, major ports in west and north Wales and the Bristol Channel. Published figures revealed that the total number of vessels sailing in and out of the Bristol Channel in 2010 was 3969, and the total number for the ports of West and North Wales combined was 6871. (Port Statistics, Department for Transport, 2013).
6.2.2 Ancillary data

AIS data are now widely utilised for shipping noise studies (e.g. Hatch et al., 2008; Merchant et al., 2012) as they provide positions of vessel movements on the sea, the vessel size and the operational speed which can be used to estimate the SL of noise generated by a ship. The realistic operational information and ship properties used in this study are extracted from an AIS database through a web-based ship tracking website (www.shipais.com). The telemetry diving data of grey seals are provided by the Sea Mammal Research Unit, the University of St Andrews and the Centre National De La Recherche Scientifique, the University of La...
Rochelle, France. The data are recorded by attaching electronic tags on marine animals, allowing tracking movements of animals and monitoring their foraging behaviours in the sea. The data also provide historic locations of seals, along with their movement speed.

6.2.3 Calculation of TL, SL, RL and SEL

The TL fields are calculated by using the coupled and the geometric spreading models respectively. Given that sound propagation is greatly dependent on the location of the source, two different transects (A and B in Fig. 6.1) are selected to represent different oceanographic and acoustic settings: transect A represents the down slope case condition with the source located in shallow mixed water while transect B represents the upslope condition with the source in the deeper stratified water. Red dots in the figure represent the locations of the sources. These two transects are used to investigate the differences of TL predicted by the coupled model and the simplified spreading models.

The solid line with travel time stamps (Fig. 6.1) gives an example of a real shipping track of a commercial cargo ship (MMSi: 353633000) in August 2010. The ship, which has a length of 508 feet and average speed of 15.5 knots, is located in the heavily used shipping lane in the Celtic Sea. The ship is modelled as a point source and the narrow band spectrum of the SL is calculated using the classic Ross (1976) power law model:

\[ L_s(f, v_s, l) = L_{s0}(f) + 60 \log\left(\frac{v_s}{12}\right) + 20 \log\left(\frac{l}{300}\right) + df \times dl + 3.0 \]  \hspace{1cm} (6.1)

where \(v_s\) is the ship’s speed in knots and \(l\) is the ship’s length in feet. \(L_{s0}\) is a reference spectrum defining an average ship as one with a speed of 12 knot and a length of 300 feet. \(df\) and \(dl\) are additional length corrections which can be found in Breeding, et al. (1996). This model is an empirical model based on a large number of measurements, which is also used by the worldwide noise models such as RANDI (Breeding, et al., 1996). Given that the
predominant bandwidth for shipping noise ranges from 10Hz to 1000Hz, the modelled narrow band spectrum from Equation 6.1 is integrated into the standard 1/3 octave central frequency spectrum between 10Hz to 1000Hz as shown in Fig. 6.2, giving the representative SL for this ship.

![Graph](image)

**Fig. 6.2** The source level spectrum in 1/3 octave band radiated by a commercial cargo ship (MMSI: 353633000).

RL is estimated using the sonar equation:

$$RL = SL - TL$$  \hspace{1cm} (6.2)

where SL is the source level in dB re 1µpa @1m predicted from equation 6.1 and TL is the transmission loss calculated by HARCAM. The TL is calculated for all 1/3 octave central frequencies between 0.01-1kHz. Accordingly, the RL at all 1/3 octave central frequencies can be obtained using this equation. The SD for all simulations is deployed at 7m, a typical depth of ship propellers (McKenna, et al., 2012). In order to simulate the horizontal spatial pattern of noise field, the multi-transect RL fields generated by the point sources A and B (see Fig. 6.1) are also calculated, with an azimuthal resolution of 2.5° and a range of 120km. For each transect the environmental conditions and sediments are fully range dependent.
SEL is a proper exposure metric of marine mammals for shipping noise (Southall, et al., 2007). It is a cumulative measure of received sound energy in time and originally defined by,

\[ SEL = 10 \log 10 \left( \frac{\int Q(t)dt}{P_0^2} \right) \]  

(6.3)

where \( Q(t) \) is the received mean square pressure and \( P_0 \) is the reference pressure of 1\( \mu \)pa. In order to calculate the SEL when this ship passes the entire track indicated by the black solid line in Fig. 6.1, an approximate discretisation is adopted here by,

\[ SEL = 10 \log 10 \left( \sum R_i T_i \right) \]  

(6.4)

Taking the spectrum of SL into account, the received mean square pressure is calculated by,

\[ R_i = \sum \frac{P_{RL}^2}{N} \]  

(6.5)

where \( N \) is the number of central 1/3 octave band frequencies between 10Hz and 1000Hz. \( P_{RL} \) is the individual received pressure derived from Equation 6.2, where the RL in decibel are converted to pressure. \( T_i \) is the duration in seconds between two time points sampled along the ship track. For this case, 1200 seconds are selected for this large geographic area to ensure sufficient resolution whilst not being overly computationally expensive.

6.3 Results

6.3.1 TL of transect A

Fig. 6.3 shows the temperature structure, the TL of transect A and the comparison of TL to the generic spreading models for summer and winter, calculated at a defined frequency of 125Hz. It is one of two 1/3-octave bands (63 and 125Hz) to be monitored proposed by MSFD. The bathymetry of transect A varies from ~30m near Land’s End to ~80m in the north. Based on the operational time of the ship extracted from the AIS database, this ship is
CHAPTER 6. SEASONAL CHANGES IN SHIPPING NOISE EXPOSURE EXPERIENCED BY DIVING SEALS

found to travel northwards through the same shipping lane (Fig. 6.1) on 05 Aug and 05 Dec in 2010. These two days are, thus, chosen for transect A, representing the summer and winter conditions respectively.

This transect is the same as in Fig. 5.4, the temperature patterns are, thus, not discussed here. The TL calculations are different to that in Fig. 5.4, with the frequency being replaced by 150Hz. The spatial patterns of TL are similar to that in Fig 5.4, but differences are still seen in Fig. 6.3c and Fig. 6.3d. In summer, the TL calculated at 150Hz is even higher (~3dB) in the upper layer than that of 300Hz (Fig. 5.4), whereas the TL is reduced slightly in winter. But this does not affect the fact that the seasonal thermocline and bottom fronts have strong impacts on the spatial pattern of the shipping noise. The vertical difference in the noise level at a frequency of 125Hz is as large as ~20dB in summer (see Fig. 6.3c). A quiet zone is, thus, formed, with the boundary following the thermocline.

Fig. 6.3e and Fig. 6.3f show the comparison of TL at a water depth of 7m predicted by the coupled model to that of the spreading models, where large differences are evident. For the summer case (Fig. 6.3e), all spreading models fail to predict the viability of TL. The TL modelled by HARCAM is between the spherical spreading and intermediate empirical spreading models to the range ~17km, subsequently following spherical spreading. The largest difference of TL is found between HARCAM and the cylindrical spreading model, reaching ~40dB, while it is less for the intermediate empirical spreading showing a typical difference of ~20dB. Spherical spreading model overestimates the TL by ~8dB at shorter range (<15km). In winter (Fig. 6.3f) the intermediate empirical spreading demonstrates skills in the prediction of TL and matches with HARCAM very well in such mixed water. TL from cylindrical spreading is lower than that of HARCAM by ~20dB whereas the spherical spreading is greater by ~20dB.
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Fig. 6.3 Seasonal variations of parameters on transect A: (a) and (b): temperature; (c) and (d): transmission loss at frequency 125Hz with source depth at 7m; Dots represent the positions of sources; (e) and (f): Comparisons of transmission loss at a water depth of 7m between HARCAM and spreading models.
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6.3.2 TL of transect B

When the ship travels northwards, the source moves to the stratified deep water. Fig. 6.4 is constructed in the same way as Fig. 6.3, but showing the TL of transect B (see Fig. 6.1). Compared with the source located in shallow mixed water (Fig. 6.3), the water column structure, bathymetry, surface wind speed and sediments are changed. The bathymetry ascends from ~110m eastwards to ~70m. The temperature fields and the associated TL patterns are very different between summer and winter. In winter the water column is well mixed, with slight horizontal variations (Fig. 6.4b). In summer strong a thermocline is developed and varies slightly in depth along the range, with larger anomalies occurring between ~21km and ~34km (Fig. 6.4a). A weak bottom front is formed at a distance of ~21km, trapping a mass of cold water (<11°C). The vertical temperature contrast is larger before ~20km (~7°C), dropping to ~5.5°C after.

In summer (Fig. 6.4c) when acoustic rays propagates through the thermocline, they are refracted into higher grazing angle ones, propagating down straightforward to the bottom. In this condition, a fraction of low angle energy is forced into higher grazing angles, thus not only increasing the number of times of interaction with the bottom but also encountering stronger bottom losses with each interaction. This leads to that the propagation range is shortened dramatically in summer, showing a difference of ~24km for the 75dB cut-off between two seasons. In the surface layer the rays are bended slightly far away from the source as the energy reflected back by the bottom is refracted again by the thermocline, leading to lower TL in the upper layer.

As seen from the TL field in winter (Fig. 6.4d), A noticeable feature can be observed that the sound rays can propagate a certain of distance in the horizontal direction before reaching the bottom. This is resulted from the positive gradient of sound speed in the vertical direction.
since sound speed increases along with the depth in such well mixed water. Such propagation pattern can reduce the interaction of rays with the bottom, and thus decreasing the bottom attenuation and loss. The TL reaches ~75dB at a range of 40km, which is similar to that in Fig. 6.3d although the location is changed. This demonstrates an insensitivity of TL to the change of the source location in winter.

**Fig. 6.4** Seasonal variations of parameters on transect B: (a) and (b): temperature; (c) and (d): transmission loss at frequency 125Hz with source depth at 7m; Dots represent the positions of sources; (e) and (f): Comparisons of transmission loss at a water depth of 60m between HARCAM and spreading models.
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By comparing the TL at a depth of 60m between spreading models and HARCAM (see Fig. 6.4f), the intermediate empirical spreading cannot capture the TL even in winter as the location differs from that in Fig. 6.3f. The TL modelled by HARCAM is ~4dB higher than that of the intermediate empirical spreading model. In summer the TL calculated by HARCAM is between that of intermediate empirical spreading and spherical spreading models (Fig. 6.4e). The largest difference (~42dB) occurs between HARCAM and cylindrical spreading model at 40km.

6.3.3 RL of sources A and B

Based on the multi-transect simulations, the spatial and temporal RL fields with radii 120km for source A are illustrated in Fig. 6.5. The RD at two water depth (15m and 50m) are plotted, representing the RL in the upper layer and bottom layer respectively. In winter, the received energy is less sensitive to the azimuthal angel, giving slight spatial variations in direction. The vertical difference is small, with the RL in the surface layer being slight higher than that of bottom water (Fig. 6.5b and Fig. 6.5d).

In summer the spatial directionality is extremely pronounced. The RL in the surface layer (Fig. 6.5a) is higher along the isobaths of northern Cornwall coast as it has an isotropic topography with lower bottom loss and well mixed water with weak downward refractions. In contrast, the sound energy decays more rapidly in the NNW directions. Large quieter surface area is formed in the deeper waters (Fig. 6.5a) when the source is deployed on the inshore side of the bottom front. Different patterns are seen in the bottom layer, showing much higher RL indicated as Fig. 6.5c. This is due to the downslope effect and bottom fronts (see Fig. 6.3), leading to a large amount of energy being trapped below the thermocline. By giving a specific RL, 100dB for instance, the difference of propagation range between upper and bottom layers reach up to ~40km. The longer the range is, the bigger the difference is.
Fig. 6.5 Radial graph of transmission losses from an omnidirectional source deployed at A (see Fig. 6.1) with frequency 125Hz and source depth 7m: (a) and (b): transmission loss at 15m; (c) and (d): transmission loss at 50m.
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6.3.4 Sound Exposure Level

The cumulative SEL calculated by equation 6.4 is a measure of sound intensity during a period when the ship travels along the shipping lane marked in Fig. 6.1. Fig. 6.7 shows the...
SEL patterns in summer and winter at two water depths (15m and 50m). The spatial patterns in summer are much more complicated than that in winter. The highest SEL exists along the ship lane and the surface layer is always louder than the bottom layer in various scales. The propagation of shipping noise is more efficient in winter than that in summer, showing higher SEL levels in winter. The magnitude of the seasonal difference increases with the distance increasing from the ship lane, reaching ~10dB near the Bristol Channel for example (see Fig. 6.7a and Fig. 6.7b). In summer the propagation range is significantly compressed and the difference in the propagation range for a specific threshold (e.g. 140dB) is ~60km shorter than that in winter. This suggests that the coastal regions have higher potential to be affected by shipping noise in winter than in summer.

In the bottom layer, it can be seen that ships in shallow mixed waters (e.g. near the Land’s End) contribute more to the bottom SEL to those in deeper waters due to the downward refraction effects (see Fig. 6.3 and Fig. 6.4). Compared with the surface layer, the seasonal difference of SEL in the bottom water is more intense, reaching ~10dB difference in a shorter range as shown by the contour lines. The spatio-temporal anomaly of SEL can potentially reflect the alteration of the radii of zones of noise influence which are the categorised ranges describing the severity of adverse effects for marine mammals (Richardson et al., 1995). This highlights strong seasonal changes of the influential range of noise, being large in winter.
Fig. 6. 7 Accumulative sound exposure level with the ship (MMSi: 353633000) travelling along the shipping lane (see Fig. 6.1).

6.3.5 Noise pattern along diving profiles

The noise exposure experienced by two seals (Seal1 and Seal2) is calculated by linking with the seal track data and the associated diving profiles. The schematic diagram (Fig. 6.8) shows the relative positions between ships and seals, the details of which are summarised in table 6.1. Seal1 corresponds to the summer condition whilst Seal2 is for the winter. Seals and ships are modelled as moving receivers and sources simultaneously. In order to compare the seasonal differences of the noise patterns resulting from environmental parameters, Winter-
ref in Table 6.1 is configured artificially the same as the case of Summer, but only with the water column data replaced by winter temperature and salinity for the TL calculation. Similarly, an artificially configured Summer-ref is used for comparison.

<table>
<thead>
<tr>
<th>Table 6.1 Summary of details of ships and seals</th>
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<tbody>
<tr>
<td><strong>Name</strong></td>
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<tr>
<td><strong>Summer</strong></td>
</tr>
<tr>
<td>Ship1 Auteuil (MMSI:563483000)</td>
</tr>
<tr>
<td>Seal1 hg29-11-10</td>
</tr>
<tr>
<td><strong>Winter-ref</strong></td>
</tr>
<tr>
<td>Ship1 Auteuil (MMSI:563483000)</td>
</tr>
<tr>
<td>Seal1 hg29-11-10</td>
</tr>
<tr>
<td><strong>Winter</strong></td>
</tr>
<tr>
<td>Ship2 Oscar Wilde (MMSI:308847000)</td>
</tr>
<tr>
<td>Seal2 B29</td>
</tr>
<tr>
<td><strong>Summer-ref</strong></td>
</tr>
<tr>
<td>Ship2 Oscar Wilde (MMSI:308847000)</td>
</tr>
<tr>
<td>Seal2 B29</td>
</tr>
</tbody>
</table>

*In order to compare to the summer condition Winter-ref is configured the same with the case of summer, but with the water column data replaced only by winter temperature and salinity. Similarly for Summer-ref.

The diving profiles and noise patterns are illustrated in Fig. 6.9 where the instantaneous RL at a frequency of 125Hz is calculated. The black arrows in the figure mark the time point when the ship crosses the bottom front from inshore to offshore while the numbers are indices of diving. As seen in Fig. 6.9a, the sound levels perceived by Seal1 are highly variable in time and water depth. Sound intensity is always low (~60dB) when the seal cruises near the sea surface (above 1.5m water depth). This is likely due to scatters of sound energy at the surface.
boundary. It is evident that the RL of every signal diving profile exhibits strong variability, showing frequent step changes of sound perceived by the seal. For instance, the change of sound level is as large as ~20dB when the seal dives from a depth of 20m to 30m (diving 28). The diving behaviour also exhibits strong difference. When the ship is located on the inshore side of the subsurface front, the seal prefers shallow diving behaviours, with only one deep diving marked by number 7 (Fig. 6.9a). The corresponding RL above a water depth of ~30m is low (70dB-85dB) whereas down to the deeper water the RL is as high as ~90dB (see diving 7). This is coincident to the case of downslope sound propagation in summer described in Fig. 6.3a. In contrast, on the offshore side of the front deep diving down to the seabed is more frequent, the sound propagation condition of which can be referred to Fig. 6.4a. The sound level is reduced when the seal dives into the bottom layer, resulting in a difference of ~15dB by comparing diving 28 to 7. This may but not necessarily indicate that the increased frequency of deep diving, when the ship moves to the deep stratified water, is caused by the reduction of noise in the deeper water depth since noise from other ships in this study is not considered. But this offers an insight that behaviour changes of seals can be linked simultaneously to noise levels by using the coupled modelling system.

Fig. 6.9b is constructed artificially the same as Fig. 6.9a, only with the water column parameters being replaced by winter temperature and salinity. It is still useful to examine how the water column properties affect the noise levels although it is not a realistic diving pattern. It can be observed that the RL difference between diving 28 and 7 in the bottom layer has disappeared. When the ship travels to the offshore side, the received sound intensity is much higher than that in summer (see Fig. 6.9a) due to lower TL. The vertical variability of RL can be still seen which cannot be explained by Fig. 6.4d. This is because the seal and ship are modelled as moving receivers and sources respectively. The propagation conditions (e.g. the
bathymetry, sediment, water column data and propagation range) are changing continuously, which extremely increase the complexity of the sound exposure of diving seals.

Fig. 6.9c illustrates the noise pattern for Seal2 (see Fig. 6.8), giving the realistic diving profiles in winter. In general the RL is much higher than that of Seal1 because the source level of Ship2 is significantly larger due to increased size and operational speed (see Table 6.1). Step changes in vertical direction can be also seen from several diving profiles, 32-40 for instance, but not as frequent as Seal1. In the case of Summer-ref (see Fig. 6.9d), step changes of sound become more frequent compared to that of winter. On the offshore side the RL is decreased remarkably (Fig. 6.9d), most likely due to the strong effects of refraction at the thermocline.

**Fig. 6.8** Schematic diagram showing the tracks of seals and ships, the details of which are summarised in Table 2. Seal1 corresponds to the summer condition whilst Seal2 is for the winter.

The accumulative SEL for Seal1 and Seal2 is shown in Fig. 6.10, giving a constant seasonal difference of ~5dB for the two seals. In summer a step change of SEL can be also observed when the seal travels from 0 to ~3km. The difference of realistic SEL between Summer (red)
and Winter (blue) reaches as much as a ~20dB, primarily due to the difference in the source level and changes of water column parameters.
CHAPTER 6. SEASONAL CHANGES IN SHIPPING NOISE EXPOSURE EXPERIENCED BY DIVING SEALS

RL(dB): 125Hz, Summer

RL(dB): 125Hz, Winter-ref

RL(dB): 125Hz, Winter
CHAPTER 6. SEASONAL CHANGES IN SHIPPING NOISE EXPOSURE EXPERIENCED BY DIVING SEALS

Fig. 6.9 Received level (dB) at frequency 125Hz by two seals (see Fig. 6.8) along the diving profiles against the relative distance between seal and ship. The black arrow in the figure marks the time point when the ship crosses the bottom front from onshore to offshore while the numbers are indices of diving. (a) Received level for Seal1 in summer; (b) The same as (a) with the water column replaced by winter temperature and salinity for transmission calculation; (c) Received level for Seal2 in winter; (d) The same as (c) with the water column replaced by summer temperature and salinity for transmission calculation.

Fig. 6.10 Cumulative sound exposure level for Fig. 6.9.
6.4 Discussion and conclusion

Underwater ambient noise generated by shipping activities has increased significantly over the past decades (e.g. Mcdonald et al., 2006; Hildebrand, 2009; Chapman and Price, 2011) and raised the concern of its various detrimental effects on marine organisms (European Commission Decision, 01/09/2010). Shipping noise is a major contributor to the low frequency noise level in coastal waters, with the dominant frequency band being 10Hz-1000Hz. Baleen whales, seals and sea lions have been found to be very sensitive to underwater shipping noise as they are low frequency users (Southall et al., 2007). Variations of shipping noise in time and space and its impacts on marine life are, however, still largely unknown as indicated by MSFD.

Sound propagation in dynamic shallow waters is extremely complicated due to high variability of oceanic processes (Katsnelson et al., 2012). In this study the impacts of stratification and associated bottom fronts on the propagation of shipping noise are shown clearly (Fig. 6.3 and Fig. 6.4). The mechanisms are explained in detail in chapter 5, and hence not discussed here. Geometric spreading models are tools to estimate sound propagation and impact zones (Richardson, et al., 1995). Results from comparisons between HARCAM and generic spreading models have shown that the cylindrical spreading model totally fails to estimate the TL in such shallow water as the difference is as large as ~20--~40dB (Fig. 6.3e), that is comparable to the field measurements in a coastal water by Pine et al. (2014). The intermediate empirical spreading is, somewhat, able to predict the TL in winter when the water is well mixed (Fig. 6.3f). The spherical spreading model which is generally used for deep waters (Jensen et al., 2011) always overestimates the TL, with a maximum value of ~20dB (Fig. 6.3e and Fig. 6.4e). It can be seen that these models are time-depth-independent, which do not take the environmental variability into account. However, sound propagation is
highly dependent on the environmental conditions in shallow waters. Consequently, the spreading models are unlikely to accurately capture the TL fields as the misfit to the coupled model (HARCAM) is large, or they may be only applicable under several limitations (e.g. for the specific location and specific time period). A study from Madsen et al. (2006) has also revealed that simple geometric models cannot be used to establish impact zones during the exposure experiments by using on-board acoustic recording tags in which the perceived sound by animals exhibits strong range and depth correlations.

The spatial distribution of noise characterised by SEL (Fig. 6.7) demonstrates strong discrepancy between summer and winter. A notable feature is that the propagation of noise is much more efficient in winter. At a defined value of SEL, the propagation distance is greater by tens of kilometres in winter than in summer. This has clear effect on the zones of noise influence (Richardson et al., 1995), which are spatial representations to classify the noise distribution based on the injury severity. It is centred at the source with the most intensive injury, lessening away from the source. The SEL characteristics predicted suggests that the range of zones determined by the severity threshold can be modified in different seasons or in the same season but different water depth. The Celtic Sea has been found to be of great ecological importance for many marine animals, including seals (Reid et al., 2003; Hammond et al., 2008). These animals may enter the exposure areas in winter but could be avoided in summer or exposed in higher injury zone in winter but lower in summer, particularly for the baleen whales and seals as they use such low frequency sound for communication (Clark, et al., 2009). So, during the spatial planning of noise zones, the seasonality of noise should be taken into account since the range difference is up to tens of kilometres in different seasons. In addition, in the designation of Marine Protection Areas (MPAs) the buffer zone is proposed as an addition to the protected zone as it is insufficient spatially to cover such
chronic exposure (Wright et al., 2011). In this condition, the radii of buffer zones could be much larger in winter but the space and cost could be reduced in summer.

Behavioural disturbance is one of many possible consequences of underwater noise, but it has great influence on energetic cost to animals for survival, hence affecting navigation, foraging and reproductive activities in marine habitats (Richardson et al., 1995). Several studies have shown that animals exposed to shipping noise change their fluke rate and dive depth (e.g. Nowacek et al., 2004), increase the amplitude of calls (Parks et al., 2011) and reduce the calling rates (e.g. Lesage et al., 1999), but research investigating the disturbance of the foraging behaviour of grey seals is limited. In this study the complex noise exposure during the diving of grey seals is shown, demonstrating strong variations in range and depth. Such variability is also observed from the controlled exposure measurements by attaching acoustic tags on animals (Madsen et al., 2006). This indicates that the coupled model used by this study has the ability to reproduce such complicated noise patterns of animals while diving. The results also show evidence that seals experience strong and frequent step changes in the sound level while diving, especially in the summer time. This is highlighted for grey seals here since they are benthic foragers, hence having higher opportunities to encounter step changes of sound and behavioural disturbances. The responses of grey seals to the step change of sound are not known yet. Verification to confirm such disturbance in behaviours may be achieved by investigating the path change and changes in diving based on the telemetry diving data.

Previous evidence (Cosens and Dueck, 1993) has shown evasive action of animals when exposed closer to a source. For the pup seal (Fig. 6.9a), changes in diving depth have been found when the ship travels from onshore to offshore side of the bottom front. When the ship is located inshore where the noise level in the bottom layer is always high (see Fig. 6.3c and
diving 7 in Fig. 6.9a), the pup seal (Seal1) prefers shallow diving. A conclusion that this pup seal tends to avoid the high noise exposure during their diving cannot be made here, since sound radiated from other ships is not considered in this study and this study is an attempt to model the likely noise exposure while foraging. The cumulative effects of shipping components have been found to be very significant to the estimation of shipping noise (Heitmeyer et al., 2004; McKenna et al., 2012). This is a subject of on-going research to include the local shipping components to study the cumulative effect on sound levels. However, by using a coupled ocean-acoustic modelling, real-time AIS data and seal diving data this study provides an approach that can be used for studies of behavioural disturbance of animals and, where possible, for other applications.

Seal2 is an adult species and its noise exposure and associated diving behaviours (Fig. 6.9c) are different from those of the pup seal. Although the overall noise level of the adult seals is much higher than that of the pup seal due to stronger source level, deep diving is still very frequent and regular for Seal2. This highlights that pup seals may be more sensitive to noise than adults.

The in-situ acoustic measurements for noised perceived by animals are rare due to a number of reasons. It is expensive and new types of acoustic tags need to be developed in order that these tags can be attached on skin of animals to record noise levels and diving parameters over a long time. Traditional measurements (e.g. Madsen et al., 2006) are focused on a few species and the noise is only recorded in a short duration, which is not enough to for the behavioural disturbance analysis. The old acoustic tags are not applicable for shipping noise as they cannot be attached for a longer time. Shipping noise is, however, a type of chronic stressor, which requires long-term noise recording. In addition, due to lack of measurements,
the modelling system cannot be fully validated. This limits the modelling tools for wider applications.

In summary, using a coupled ocean-acoustic modelling system by compiling water column data, sediments, bathymetry and sea surface wind speed, significant impacts of oceanic features on shipping noise propagation have been shown, leading to strong spatio-temporal variations in the predicted noise patterns. In summer the areas of high noise exposure are situated below the thermocline when the ship is located on the inshore side of oceanic fronts, and above the thermocline when the ship is on the offshore side of oceanic fronts. The difference in sound level between the top and bottom of the water column is as high as ~20dB. Shipping noise propagated much further (tens of kilometres) in winter than in summer. The cumulative SEL experienced by seals is persistently higher by ~5dB in winter than in summer. Comparisons of TL modelled by the coupled model to that of geometric spreading models have been carried out where differences were found up to ~40dB in summer. By overlapping the realistic AIS shipping track data and the telemetric seal diving data, the noise exposure experienced by diving seals have been estimated in summer and winter. The results show that noise patterns perceived by seals are extremely complicated, demonstrating high dependence on the time and water depth. Frequent step changes in the received sound level were seen during their descent/ascent through the water column.

Although the chronic effects on marine life or ecosystem from shipping noise is still largely unknown and have not been incorporated into management decisions (Ellison, et al., 2012), there is increasing concern about such detrimental impacts on animals. Changes in environmental conditions can lead to significant variations of sound propagation, thus increasing the uncertainty in the acoustically sensitive area. Compiling the detailed environmental conditions into shipping noise management is beneficial not only to
CHAPTER 6. SEASONAL CHANGES IN SHIPPING NOISE EXPOSURE EXPERIENCED BY DIVING SEALS

understand how the dynamic acoustic environment affects the noise field, and thus the marine organisms, but also optimise the noise mitigation strategy and the designation of MPAs. Such coupled model, as a cheap but reliable method, can simulate the sound field over large area and long period. It is important for either biological application or anthropogenic noise estimation, especially for the continuous acoustic events in a big geographic area (e.g. shipping noise) as a complete measure is impractical or expensive.
Chapter 7

7 Summary and future study

7.1 Summary of conclusions

A baroclinic three-dimensional primitive equation finite difference model (POLCOMS) has been applied in the Eastern Celtic Sea in order to simulate the thermodynamics. The ability of the optimally configured POLCOMS model to resolve vertical thermal structures is assessed by comparison of model simulations with observational data collected in the Celtic Sea in 1998, giving good model skills.

Sensitivity of model outputs to atmospheric forcing taken from different sources is investigated, showing significant differences between modelled results. First, the study examines if the performance of an ocean model is improved when the data to force the model is taken from a higher resolution atmospheric model obtained from the British Atmospheric Data Centre (BADC, 0.11° resolution), while the coarser resolution data is extracted from NCEP-II with 1.6° resolution. The predicted SST from BADC is, however, continuously lower through the year than the observed, reaching a maximum of approximately -1.15°C in August after the period of strongest solar radiation in July. It is caused primarily by the much denser total cloud cover provided by BADC, which significantly reduces the downwelling solar radiation. As the resolution is an important factor to the choice of surface forcing, one might prefer with higher resolution forcing to drive the ocean model. However, the results indicate that high resolution data does not necessarily promote the model accuracy. The
inherent quality of forcing may be of the most importance to the model predictability. Consequently, resolution should not be treated as an isolated factor when selecting surface forcing.

Model results forced by five popular reanalysis products are compared in order to examine how sensitive the modelled results are to the selection of surface forcing. The results reveal that all products have analogous capability in predicting the spatially averaged SST, but the water column structures. By comparing to the observations, NCEP predicts the most accurate thermal structures while the depth of thermocline from models driven by JRA25 and ERA40 is quite shallow, owing to the weaker wind speed. It is worth noting that JRA55 and ERA-interim exhibit improvements compared with JRA25 and ERA40 since they are the upgraded versions. The seasonally frontal position is insensitive to surface forcing whereas the strength of stratification, the depth of thermocline and the temperature contrast at the thermocline demonstrate high dissimilarity to different forcing, with the most marked differences occurring in summer.

The variability of TL, varying from seasonal to hourly scales, is studied using the coupled ocean-acoustic model, showing strong dependence on the geographic locations of sound source and its depth. In summer, when the source of sound is on the inshore side of the bottom front, the sound energy is mostly concentrated in the near-bottom layer (between seabed and 20m level). In winter, the sound from the same source is distributed more evenly in the vertical. When the source of sound is on the seaward side of the front, the sound level from shallow source is nearly uniform in the vertical and the transmission loss is significantly greater (~16dB at 40km distance) in the summer than in the winter. In contrast, sound energy from deep sources is trapped in the bottom cold water, leading to a much lower transmission loss (~20dB) in summer than in winter, or ~10dB fluctuation during the deterioration of the
thermocline in late autumn. The shallow sources (e.g. ships) are sensitive to the surface heat flux as it changes the vertical temperature gradient, hence affecting the refraction of rays at the interface of thermocline. Tides play an important role for deeper sources (e.g. pile driving) in determining the TL variability since it changes the position of subsurface fronts.

Shipping noise is a major contributor to anthropogenic noise in the sea, which is now classed as pollution in accordance with the MSFD. However, little is known about how it impacts marine organisms. This study investigates potential shipping noise experienced by grey seals (*Halichoerus grypus*) in the Celtic Sea by overlaying their GPS tracks and dive data, over a state-of-the-art ocean (POLCOMS) and acoustic (HARCAM) propagation model populated with real-time AIS shipping data in summer and winter. The results show a clear influence of the seasonal thermocline on shipping noise propagation. In summer the areas of high noise exposure are situated below the thermocline when the ship is located on the inshore side of oceanic fronts, and above the thermocline when the ship is on the offshore side of oceanic fronts. The difference in sound level between the top and bottom of the water column is as high as ~20dB. Shipping noise propagated much further (tens of kilometres) in winter than in summer. The cumulative SEL experienced by seals is persistently higher by ~5dB in winter than in summer. Furthermore, this study shows strong step changes of sound perceived by seals during their descent/ascent through the water column. Since grey seals tend to be benthic foragers, the step-change in sound exposure may have negative impacts on their foraging behaviour. It is only through a more realistic understanding of exposure of animals to ship noise that we can set appropriate management and mitigation targets.

### 7.2 Future study

As the propagation loss is crucially dependent on the 3D structure of the water column, it should be supplied to HARCAM at high horizontal and vertical resolution. The development
of POLCOMS has been ceased and continuous efforts have been made and shifted to the development of NEMO. Consequently, computations of the temperature and salinity will be carried out using a modern 3D ocean model (NEMO-Shelf), which is a state-of-the-art modelling framework for oceanographic research, operational oceanography seasonal forecast and climate studies (Madec et al., 2012). It is being continuously improved and used by a large community. The University of Plymouth contributes to the development of NEMO-Shelf as a partner in EU MyOcean and PERSEUS projects and has a track record of using NEMO-shelf for research purposes (e.g. Shapiro et al., 2013; Wobus et al., 2013). The ocean model will be configured using the latest development by the NEMO community (Madec et al., 2012) including the hybrid enveloped SZHE vertical discretisation (Shapiro et al., 2013), the Smagorinsky parameterisation for horizontal viscosity and diffusivity and the non-linear free surface with vertical piecewise parabolic method (vPPM) for vertical advection.

In this study a single ship track is used to estimate the noise pattern in order to emphasise the effects of environment conditions on sound propagation. The shipping component is, however, an important factor to affect the noise distribution (Heitmeyer et al., 2004; Merchant et al., 2012). Models to include multiple sources (ships) are indispensable for mapping the general noise pattern in coastal waters. In the future studies, all ships will be taken into account to predict the more realistic noise intensity in the sea by using the latest AIS database, especially to monitor noise levels at two 1/3-octave bands (63 and 125Hz) over the year as proposed by MSFD.

The impact of sound step change on seals is suggested by this study as a hypothesis, tests are, therefore, essential to verify this question. Responses to increased sound levels will be investigated by examining path change and changes in diving behaviour. The region will
include routes travelled by seals between Brittany, Cornwall and South Wales between their key breeding and foraging sites and tracks of animals traversing extremely busy shipping lanes in the English Channel and the Celtic Sea where large archived diving data are available. The seal telemetry data will be supplied by colleagues at the Sea Mammal Research Unit at the University of St Andrews, CNRS, University of La Rochelle, France and University College Cork, Ireland and will include tracks of seals travelling and foraging in an area around the SW of England. These data will be used to examine the marine mammal movement and diving behaviours. By integrating the acoustic noise into behavioural patterns of marine fauna, the acoustic impact from ships will be studied from individual to population levels of animals. Only through a more realistic understanding of animals to ship noise can we understand the level of disturbance due to acoustic pollution and set appropriate management and mitigation targets.

Before applying the modelling system to a wider application, the uncertainties of the system should be investigated to make it robust and reliable. The shipping AIS data should be examined since the dataset consisting of operational information and ship properties does not include completely all records of ships. It is only compulsory for vessels exceeding 299 gross tonnes to install the AIS transceivers in accordance with the international convention for the Safety Of Life At Sea (SOLAS, 2005). Vessels below 299 gross tonnes which have been increasing in the number are not included in the database, which potentially underestimates the source levels in oceans. These vessels also have different types (e.g. fishing boats, recreations boats and research vessels), hence different noise spectrums. The representative source level for each type of vessels could be obtained by using the statistical method introduced by Erbe et al. (2012). The coupled ocean-acoustic modelling system should be validated against observations. Previous validations for acoustic modelling uses static hydrophone to receive the sound energy propagated from a source. However, for behavioural
disturbance analysis of animals, the animal is a dynamic receiver and its travelling path is uncontrollable. The wave guide or the propagation condition between the source (a ship) and the receiver (a seal) changes persistently. New acoustic tags which not only record the sound energy perceived by the animals, but also include the diving parameters could be very suitable for investigating the impacts of noise on animals. It substantially links the noise exposure and diving behaviours of animals into a single device. Apart from predicting noise from shipping, the ambient noise (e.g. surface wave sound) around animals should be considered at the same time because the ambient noise might introduce slight disturbance as well.
Appendixes

A1: POLCOMS description

POLCOMS is formulated in spherical polar sigma coordinates system in horizontal with $\chi$ (eastwards) and $\phi$ (northwards), and in vertical with $\sigma = (z - \zeta)/(h + \zeta)$, where $z$ is the vertical Cartesian coordinates, $h$ is the water depth relative to the reference sea surface level ($z = 0$) and $\zeta$ is the surface elevation. The vertical coordinates $\sigma$ is split into terrain-following levels, allowing to varying in the horizontal plane. It follows the transformation of $s$ coordinates described by Song and Haidvogel (1994) and is defined by:

$$
\begin{align*}
\sigma &= s + \frac{h - h_c}{h} (C(s) - s) \quad h > h_c \\
&= s \quad h \leq h_c 
\end{align*}
$$

where

$$
C(s) = (1 - b) \frac{\sinh(\theta s)}{\sinh \theta} + b \frac{\tanh[\theta(s+0.5)] - \tanh(0.5 \theta)}{2 \tanh 0.5 \theta}
$$

and $s$ represents the $N$ evenly spaced levels between $\sigma = -(1 + 0.5(N - 2)^{-1})$ and $\sigma = 0.5(N - 2)^{-1}$. $h_c$ is the critical water depth. The deviation from the usual $\sigma$ level is given by a set of curves $C(s)$. $\theta$ and $b$ are the surface and bottom control parameters respectively. The vertical resolution allows to be increased near surface and bottom layers by changing $\theta$ and $b$.

Horizontally, POLCOMS uses the B-grid. The $u$ and $v$ components are defined at $u$-points, half a grid box to the southwest of $b$-points where elevations $\zeta$ and other scalar variables are defined. Such grid is chosen since it conserves the large horizontal density variations, hence
preserving frontal features (James, 1986). This is because the B grid does not require averaging of the Coriolis force over the grid box which generally occurs using the C grid.

**Equation of state**

The density is defined by an approximation to the full UNESCO equation \( \rho(T,S,p) = \rho(T,S,0) + \rho'(T,S,p) \), where \( \rho(T,S,0) \) is taken from the UNESCO equation of state and

\[
\rho'(T,S,p) = 10^4 \frac{p}{c^2} \left( 1 - 0.2 \frac{p}{c^2} \right)
\]

where \( c = 1449.2 + 1.34(S - 35) + 4.55T - 0.0457^2 + 0.00821p + 15.0 \times 10^{-9} p^2 \)

In order to improve the numerical calculations, a term for the variations of compressibility with temperature and salinity is included in the buoyancy equation of state (see Holt and James, 2001).

**Equation of motion**

POLCOMS solves the primitive motion equations with the incompressible, hydrostatic and Boussinesq approximations. The solution is split into a fast barotropic component (depth mean) and a slow baroclinic component (depth varying). The eastward velocity is expressed as \( u = \bar{u}(\chi, \phi, t) + u_r(\chi, \phi, \sigma, t) \) and the northward velocity \( v = \bar{v} + v_r \).

The depth mean equations are expressed by:

\[
\frac{\partial \bar{v}}{\partial t} = f \bar{v} - R \cos \phi \gamma^{-1} \left[ g \frac{\partial \zeta}{\partial \chi} + \rho_0^{-1} \frac{\partial \rho_a}{\partial \chi} \right] + H^{-1} [F_S - F_B] + NLB_\chi
\]

\[
\frac{\partial \bar{u}}{\partial t} = -f \bar{u} + R^{-1} \left[ g \frac{\partial \chi}{\partial \phi} + \rho_0^{-1} \frac{\partial \rho_a}{\partial \phi} \right] + H^{-1} [G_S - G_B] + NLB_\phi
\]
The depth varying components are defined as:

$$\frac{\partial u_r}{\partial t} = -L(u) + f u_r + \frac{uv \tan \phi}{R} - \Pi_\chi + D(u) - H^{-1}[F_S - F_B] - NLB_\chi$$  \hspace{1cm} (A1.6)

$$\frac{\partial v_r}{\partial t} = -L(v) - f u_r - \frac{u^2 \tan \phi}{R} - \Pi_\phi + D(v) - H^{-1}[G_S - G_B] - NLB_\phi$$  \hspace{1cm} (A1.7)

where $R$ is the radius of the earth and the buoyancy terms are given by:

$$\Pi_\chi = (R \cos \phi)^{-1} \left. \frac{\partial \psi}{\partial \chi} \right|_z \quad \Pi_\phi = R^{-1} \left. \frac{\partial \psi}{\partial \phi} \right|_z$$  \hspace{1cm} (A1.8)

As can be seen from equation A1.8, it has not been transformed to the $\sigma$ coordinates in order to avoid errors of calculating the horizontal pressure gradient with steep topography. A limitation of $\sigma$ coordinates is that the horizontal pressure gradient cannot be represented in the region of steep topography because the gradient is defined perpendicular to the $z$ direction. It can leads to large truncation errors of estimating the horizontal pressure gradient (Haney, 1991). Instead, the buoyancy contribution to the horizontal pressure gradient in POLCOMS is calculated by interpolating the buoyancy field onto a horizontal plan which has a $u$ grid as the centre. Improvements have been found when the thermocline is flat or sloping in the opposite direction to the $\sigma$ levels (Holt and James, 2001).

The depth means of the nonlinear and buoyancy terms are:

$$NLB_\chi = \int_{-1}^{0} \left[ -L(u) + f u_r + \frac{uv \tan \phi}{R} - \Pi_\chi \right] d\sigma$$  \hspace{1cm} (A1.9)

$$NLB_\phi = \int_{-1}^{0} \left[ -L(v) + \frac{u^2 \tan \phi}{R} - \Pi_\phi \right] d\sigma$$  \hspace{1cm} (A1.10)

The advection terms are given by:
\[ (a) = \frac{u}{R \cos \phi} \frac{\partial a}{\partial x} + \frac{v}{R} \frac{\partial a}{\partial \phi} + \Omega \frac{\partial a}{\partial \sigma} \quad (A1.11) \]

with

\[ \Omega = \frac{\sigma}{H} \frac{\partial \xi}{\partial t} - (HR \cos \phi)^{-1} \times \left[ \frac{\partial}{\partial x} (H \int_0^\sigma ud\sigma) + \frac{\partial}{\partial \phi} (H \cos \phi \int_0^\sigma v d\sigma) \right] \quad (A1.12) \]

The vertical gradient of stresses is replaced by a diffusion term as many other models used:

\[ D(a) = H^{-2} \frac{\partial}{\partial \sigma} \left( K_z \frac{\partial a}{\partial \sigma} \right) \quad (A1.13) \]

where \( K_z \) is the eddy viscosity estimated from the turbulence model.

The non-linear free surface is calculated using:

\[ \frac{\partial \zeta}{\partial t} = (R \cos \phi)^{-1} \left[ \frac{\partial}{\partial x} (H \bar{u}) + \frac{\partial}{\partial \phi} (H \cos \phi \bar{v}) \right] \quad (A1.14) \]

For the vertical boundaries, POLCOMS uses a slip boundary condition and the surface and bottom stress components and the associated coefficients are given by

\[ (F_s, G_s) = c_s \frac{\rho A}{\rho_0} (u_w, v_w) \sqrt{u_w^2 + v_w^2} \]

\[ c_s = 0.63 + 0.66 \sqrt{u_w^2 + v_w^2} \quad (A1.15) \]

Where \((u_w, v_w)\) is the wind velocity at 10 m and the bottom component is:

\[ (F_B, G_B) = c_B (u_B, v_B) \sqrt{u_B^2 + v_B^2} \]

\[ c_B = \left( \kappa^{-1} \log \left( \frac{\delta}{z_0} \right) \right)^{-2}, \quad c_B > 0.005 \quad (A1.16) \]
The near bed velocity \((u_B, v_B)\) is defined at a depth \(\delta\) above the sea bed and the roughness length is taken to be \(z_0=0.003\) m and \(\kappa=0.41\) is von Karman’s constant.

**Advection**

In order to minimise numerical diffusion and maintain horizontal gradients the Piecewise Parabolic Method (Collella and Woodward, 1984; James, 1996) is adopted to deal with the advection of momentum and scalars. This scheme assumes that variables starting with the box averaged value \(\bar{a}\) vary parabolically crossing the grid boxes through:

\[
a(\eta) = a_L + \eta[a_R - a_L + a_6(1 - \eta)]
\]

\[
a_6 = 6\bar{a} - 3(a_R + a_L) \tag{A1.17}
\]

where the coordinate \(\eta\) crosses the grid, varying from 0 to 1. \(a_L\) and \(a_R\) are left and right values at the edges of box respectively, which are found by interpolating a quartic polynomial fit to the integral of \(a\). The advective flux is calculated by integrating the parabolas using an upwind sense. This method has been found well to conserve the frontal structures (James, 1996).

**Horizontal diffusion**

POLCOMS was initially designed without the horizontal diffusion term and baroclinic eddies produced by this model in the stratified water (e.g. the Celtic Sea) in summer has been found to be too energetic, long-lived and frequent. The inclusion of a horizontal diffusion to the force term of the momentum and transport equations can significantly reduce the eddy kinetic energy (Holt and James, 2006). Eddies are resolved more accurately without affecting significantly the results in other regions by adding such horizontal diffusion term.
The formulation of diffusion for momentum and transport used in the model is the Laplacian diffusion term:

\[
F_m = \nabla \cdot A_m \nabla \mathbf{u} \quad F_T = \frac{1}{Pr} \nabla \cdot A_m \nabla \phi
\]  

(A1.18)

Where \( Pr \) is the turbulent Prandtl number and \( A_m \) is the horizontal turbulent diffusivity calculated using (Smagorinsky, 1963)

\[
A_m = \alpha \Delta x \Delta y \left[ \left( \frac{\partial u}{\partial x} \right)^2 + \left( \frac{\partial v}{\partial y} \right)^2 + \frac{1}{2} \left( \frac{\partial u}{\partial y} + \frac{\partial v}{\partial x} \right)^2 \right]^{1/2}
\]  

(A1.19)

where \( \alpha \) is a constant and \( \alpha = 0.2 \) has been suggested to be an appropriate value for a 1.8 km model in the Celtic Sea (Holt and James, 2006).

**Turbulence closure model**

The original turbulence closure model implemented in POLCOMS is the Mellor-Yamada-Galperin scheme (Mellor and Yamada, 1974; Galperin et al., 1998), which is a one equation, second moment closure model. The rate of dissipation \( \varepsilon \) is evaluated from the cascading relation:

\[
\varepsilon \propto \frac{k^{3/2}}{l}
\]  

(A1.20)

Where \( k \) is turbulence kinetic energy and \( l \) is the algebraic length scale using an asymmetric profile following Bakhmetev (1932): \( l = \kappa H(1 + \sigma)(-\sigma)^{0.5} \). \( \kappa \) is the von Karman’s constant, \( H \) is total water depth and \( \sigma \) is the vertical coordinates.

The turbulence closure model is subsequently replaced by the General Ocean Turbulence Model (GOTM; Umlauf and Burchard, 2005) through which the accuracy of modelling tidal...
mixing fronts and seasonal stratification has been promoted (Holt and Umlauf, 2008). GOTM is a one-dimensional water column model and allows various models and methods to be selected. For this study we use the two-equation and second-order model with dynamic equation for turbulence kinetic energy \((k-\varepsilon)\) type and with a dynamic dissipation rate for the mixing length, which is a length scale describing the largest eddies in a system. It is estimated using the relation:

\[
L = (c_H^0) \frac{k^{2/3}}{\varepsilon}
\]  

(A1.21)

where \(c_H^0\) is the stability function.

The model parameters for this \(k-\varepsilon\) configuration model are listed in Table A1.1.
Table A1. 1 Parameters for the k-ε scheme of turbulence closure model

<table>
<thead>
<tr>
<th>Name</th>
<th>Variable</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Prandtl number</td>
<td>$P_r$</td>
<td>0.74</td>
</tr>
<tr>
<td>Coefficient in dissipation equation</td>
<td>$c_1$</td>
<td>1.44</td>
</tr>
<tr>
<td>Coefficient in dissipation equation</td>
<td>$c_2$</td>
<td>1.92</td>
</tr>
<tr>
<td>Coefficient for stable stratification</td>
<td>$c_3^-$</td>
<td>-0.4</td>
</tr>
<tr>
<td>Coefficient for unstable stratification</td>
<td>$c_3^+$</td>
<td>1</td>
</tr>
<tr>
<td>Schmidt number for TKE diffusivity</td>
<td>$\sigma_k$</td>
<td>1</td>
</tr>
<tr>
<td>Schmidt number for dissipation diffusivity</td>
<td>$\sigma_e$</td>
<td>1.3</td>
</tr>
<tr>
<td>The desired steady-state Richardson number</td>
<td>$R_i$</td>
<td>0.25</td>
</tr>
<tr>
<td>Coefficient for length scale limitation</td>
<td>$c_{lim}$</td>
<td>0.267</td>
</tr>
<tr>
<td>Minimum TKE</td>
<td>$k_{min}$</td>
<td>$1.0 \times 10^{-6} m^{-2} s^{-2}$</td>
</tr>
<tr>
<td>Minimum dissipation rate</td>
<td>$\varepsilon_{min}$</td>
<td>$1.0 \times 10^{-12} m^{-2} s^{-3}$</td>
</tr>
<tr>
<td>Minimum buoyancy variance</td>
<td>$k_{b min}$</td>
<td>$1.0 \times 10^{-10}$</td>
</tr>
<tr>
<td>Minimum buoyancy variance destruction rate</td>
<td>$\varepsilon_{b min}$</td>
<td>$1.0 \times 10^{-14}$</td>
</tr>
</tbody>
</table>
A2: Model set-up of POLCOMS

Defining the model domain

This study focuses on an area in the Celtic Sea between 50.0750°N to 51.8333°N and 7.9°W to 4.0°W, surrounded by the north Cornish coast, the Bristol Channel, the south St. George’s Channel and the west and south opening boundary. Fig. 1.1 shows the model domain as covered by the red rectangular box. It is necessary to point out that the water located at the south-east corner of the red box (south Cornwall) will be treated as land it is not the region of the Celtic Sea. Consequently, the actual water is the region indicated by the red border in the rectangular box. In order to exclude wetting and drying effects, numerical calculations are performed only in the areas deeper than 5 m.

Defining the grids

Horizontally, POLCOMS is formulated with the Arakawa (1972) B-grid. The $u$ and $v$ components are defined at u-points, half a grid box to the southwest of b-points where elevations $\zeta$ and other scalar variables are defined (Fig. A2.1). The model domain formulated on Arakawa B-grid ranges between 50.075°N to 51.8333°N and 7.9°W to 4°W, with a horizontal resolution of 1/35° in the longitudinal direction and 1/55.735° in the latitudinal direction, respectively. The properties of the horizontal grids including the dimension of grids, the resolution, start points etc. are summarised in Table A2.1.

As discussed in the previous section, the water at the south-east corner of the study domain (south Cornwall) will be treated as land. A mask file (named as MasC.dat) is created for separating the water and land, with a value of 1 indicating the water greater than 5 m and 0 representing land areas.
Table A2.1 Summary of properties of horizontal discretization for the model domain

<table>
<thead>
<tr>
<th></th>
<th>Western limit</th>
<th>Southern limit</th>
<th>Resolution Y direction</th>
<th>Resolution X direction</th>
<th>Approx dx and dy (km)</th>
<th>Number of grids</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>u-points</strong></td>
<td>7.9144°W</td>
<td>50.0659°N</td>
<td>1/55.735°</td>
<td>1/35°</td>
<td>2 km</td>
<td>Lon: 137</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Lat: 98</td>
</tr>
<tr>
<td><strong>b-points</strong></td>
<td>7.9°W</td>
<td>50.075°N</td>
<td>1/55.735°</td>
<td>1/35°</td>
<td>2 km</td>
<td>Lon: 137</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Lat: 98</td>
</tr>
</tbody>
</table>

Fig. A2.1 Diagram illustrating the arrangement of u-points and b-points of the model domain. Blue circle points represent the b-points where the elevations and other scalar variables are defined. Red crosses indicate the u-points where the u and v components are defined.

Vertically, the water column is discretised into 30 σ levels based on the transformation of equation A1.1 and A1.2 described in Appendix 1, with parameters $h_c = 200$, $\theta = 8$ and $b=$
0.05. Fig. A2.2 illustrates the vertical discretisation of model domain, extracted from a cross section along 51.3257°N. Given that the topography in the model domain is quite shallow, the critical depth is selected to be \( h_c = 200 \text{m} \) which is greater than the maximum depth (~120m) of the model domain. The vertical discretisation is, therefore, the standard \( \sigma \) coordinates without any resolution enhancement near the surface and bottom layers although it is allowed, because the standard \( \sigma \) coordinates utilised here give adequate resolution for such shallow water to solve the oceanic processes.

Fig. A2.2 Vertical discretisation of study domain for the Celtic Sea along 51.3257°N

The bathymetry data of the Celtic Sea was acquired from the ETOPO2v2 (2006) Global Gridded 2-minute Database of the National Geophysical Data Center (NGDC). The Grid translator of Geophysical Data System (GEODAS) with friendly interface provides a convenient way to download the bathymetry data, which has a resolution of 2 minutes in both
latitude and longitude directions corresponding to latitudinal ~4km and longitudinal ~2km respectively. The downloaded bathymetry data between 49°N to 53°N and 9°W to 3°W was stored as a XYZ format. Surfer®, a software for data processing, was adopted to interpolate the 2-minute ETOPO2v2 data to the model grids defined in the previous section. The method used for interpolating and smoothing was ‘moving average’, which takes the average of all points in a defined search ellipse. In this study the value 0.1 was specified for both the semi-major and semi-minor axis. Meanwhile, a filter was defined to eliminate any data shallower than 5 m.

By coupling with the mask file (MasC.dat), MATLAB® was used to interpolate the bathymetry data onto the model grids. A 3 x 3 matrix filter was also applied to smooth the bathymetry data in order to minimise the numerical errors. The resulting bathymetry file was
then used as the input file of model runs and for calculating the levels of $\sigma$ coordinates. The final bathymetry used for model simulations is shown in Fig. A2.3.

**Defining initial temperature and salinity**

The scalar fields, temperature and salinity for example, can be either initialised by specifying a constant value in the control file `parameters.dat` or read in from an external data file. The data from data files must be three dimensional and consistent to the levels of $\sigma$ coordinates. The monthly climatological mean temperature and salinity from the World Ocean Database (hereafter referred to Q01_M) on a $1/4^\circ$ grid was chosen since it has an increased spatial resolution and reduced noise by additional smoothing in horizontal space (Boyer *et al.*, 2005), compared with the previous version. Q01_M using the dataset of World Ocean Database 2001 (WOD01), covers the world ocean for monthly periods down to a depth of 1500m with 24 stand levels. The input file of temperature and salinity conditioning to POLCOMS was created using following methods:

The monthly files were downloaded from the National Oceanographic Data Center (NODC). A FORTRAN® file initially written by D. Aleynik was used to cut the initial data from the world field into a sub-field that covers enough of the study domain. The cropped data were gridded regularly ($1/4^\circ$), ranging between 47.125°N to 55.125°N and 9.875°W to 0.125°E, with a horizontal dimension of 41 points in the longitudinal direction and 33 points in the latitudinal direction. In the vertical direction, the cropped data cover all depth levels from 1 to 24.

By means of the bathymetry and model grids produced in the previous section, the depths in $z$-coordinates can be calculated. The method used to calculate the depths of each $\sigma$ levels is illustrated as follows: the locations of $\sigma$-level (always between 0 and -1) are calculated by
coupling the depths in z-level and equation A1.1 and A1.2 in Appendix 1 of σ-coordinates. Then the value of the location is multiplied by the total depth to find the σ-level depth for the corresponding grid point in the vertical direction.

FORTRAN® was then used to interpolate the data from cropped grids to model grids. The method utilised for interpolation is described as below:

The grids marked as dots (Fig. A2.4a) represent the cropped Q01_M mesh with dimension of 41x33 while the red crosses describe the model grid with dimension of 137x98. It can be seen clearly that every red point is surrounded by 4 blue points. For clarity, a single unit for the interpolation taken from Fig. A2.4a is shown in Fig. A2.4b. A, B, C and D express the values of temperature or salinity of Q01_M and O represents the model grid to be interpolated. S1, S2, S3 and S4 are areas of four rectangles. The total area of the blue box is denoted by S. The calculation of linear interpolation is carried out using Equation A2.1.

\[
O = A \times \frac{S4}{S} + B \times \frac{S3}{S} + C \times \frac{S2}{S} + D \times \frac{S1}{S} \quad \text{(A2.1)}
\]

All values of red points in the horizontal plan can be calculated using this method above. Applying this procedure to all depths of Q01_M between 1 and 24, results in a 3D data set with dimensions of 137x98x24. In the vertical direction, coupling the 3D data set generated and depths of σ-level produced in step 2, the data is interpolated onto 30 σ layers in the vertical direction.

The monthly averaged data are designed to represent the temperature and salinity fields on the 15th of every month. As the model simulations start on the 1st of a month, in order to match each other in time, the temperature and salinity, therefore, is centred linearly to the 1st between two adjacent months. The final monthly temperature and salinity files representing
the values of the 1st for each month were read in by POLCOMS when model time reached 00:00 of the first day of a month.

Fig. A2. 4 Diagram (a) represents the grid of Q01_M with dots and model grid with crosses. (b) denotes the arrangement of interpolation. A, B, C and D express the values of temperature or salinity of Q01_M and O represents the value to be interpolated. S1, S2, S3 and S4 are areas of four rectangles respectively.

**Defining open boundary temperature and salinity**

In order to avoid small scale variations in the boundary zone, the temperature and salinity are relaxed to climatological values in a region next to model boundaries. A relaxation zone with a width of four grid points was used by POLCOMS. The locations of open boundaries can be found in Fig. A2.5, indicated by the solid black lines. The original routine (*bost.for*) for reading in boundary data could be only used for short runs as it reads in boundary values of the last month, the current month and the next month, and then interpolated to the current model time. In order to extend the simulation, O’Neill (2008) developed a new algorithm and rewrote the routine (*bost.for*) to *bostlong.for* so that it can read in boundary data over a long period. The arrangement of new boundary indices is shown as Fig. A2.5. POLCOMS reads in boundary data starting from the south-east corner and travels in a clockwise spiral from the outer circle to the inner one. The monthly boundary data, which was extracted from monthly
climatological temperature and salinity files described in the previous section, was fed into the boundary points starting from number 1 to the end point along such clockwise spirals. As it is necessary to provide POLCOMS with temperature and salinity data for the boundary zone through entire length of the run, the current month data and the next month data were interpolated linearly to the current model time. In order to deal with interpolation in time correctly, a file (boundary_TS_time.txt) containing the number of days since the start of a simulation was created to control the timing of reading boundary data of the following month.

![Schematic of the index arrangement of 4 grids width boundaries](image)

**Fig. A2. 5** Schematic of the index arrangement of 4 grids width boundaries. It starts at number 1 and continues around in a clockwise spiral. Blue circle points indicate the boundary points.

**Defining surface forcing**

A complete simulation requires surface fluxes of heat, momentum, and freshwater. Four atmospheric variables, namely mean sea level pressure, air temperature, relative humidity and total cloud cover are used to calculate the incoming solar radiation and outgoing long-wave radiation. The momentum is determined by the surface wind forcing while the freshwater is governed by the precipitation rate. A total of seven surface forcing components interpreted above are utilized for this study. One of the main focuses of this study is to investigate the
sensitivity of model predictions to different surface forcing, which were taken from different sources. Two sets of atmospheric forcing were used for the sensitivity study (see section 4.2): (i) data from the British Atmospheric Data (BADC, 0.11° resolution) and (ii) data from the National Centers for Environmental Prediction (NCEP-II, approximately 1.6° resolution), both of which have the same frequency of 6-hour interval.

NCEP-II is an improved version of the NCEP Reanalysis I model that fixed errors and updated parameterisations of physical processes. 6-hourly NCEP-II meteorological data were downloaded from NOAA PSD (NOAA/OAR/ESRL PSD, Boulder, Colorado, USA). The data was available as NetCDF format and the CSIRO NetCDF MATLAB interface was used to extract the data. Subsequently the data was interpolated from the original grids with 1.6° × 1.6° into a sub-domain of 1° × 1° covering the study area. The air temperature was originally in K and converted to °C before producing the final input files. The precipitation rate with unit kgm⁻²s⁻¹ was accumulated within day and converted into kgm⁻²day⁻¹.

Higher resolution BADC data produced by the Met Office North Atlantic European (NAE) unified model (UM) for the operational weather prediction purpose was acquired from the British Atmospheric Data Centre (BADC) under the licensed agreement with the National Environmental Research Council (NERC) and the UK Meteorological Office (UKMO). The procedures for data processing are described as follows:

a) Download NAE model data

As the datasets have restricted access due to their proprietary nature, an application of permission is necessary before using the data. The outputs from the UM model had been divided spatially into global data, mesoscale data and NAE data. The NAE data, covering years between 2006 and 2010 with a good spatially coverage over the Celtic sea, was considered for this study. The NAE datasets were stored as that for each time step (4 times
per day) there is a corresponding single file available. Consequently, it is not realistic to download all files manually. A MATLAB® script automating the downloading task was developed to connect the FTP server and download the data of interest meanwhile a text file was generated to record the information about data gaps.

b) Convert the format of data

The NAE data are binary files in the ‘*.PP’ format. Two packages (Xconv® and Convsh®) provided by the Met office can be used to read the ‘PP’ files and convert the format to a suitable one, e.g. NetCDF, which could be manipulated by MATLAB®. Xconv® is a windows-based package designed to manipulate the ‘PP’ files. Although this package with a user-friendly interface is easy to use, it can only process a few files every time due to the limit of memory. Another software (Convsh®), having the same functions with Xconv®, is a command line package, allowing scripts written by TCL language to automate various Xconv® tasks in terms of reading data, writing data and converting the format. By means of Convsh®, a TCL script was developed with the utility that it can convert the ‘PP’ format into NetCDF and compress all files to a single one as a whole.

c) Generate the NAE model coordinates

The UM model uses a rotated latitude and longitude non-standard polar axis system centred in UK, aiming at obtaining a fairly uniform horizontal resolution (approximately 12km) over the area of interest. A FORTRAN® subroutine obtained from Met Office under the agreement was used to carry out the generation of grids. The coordinates generated contain the actual latitude and longitude of every model grid (NAE model) where the data is located.

d) Extract NetCDF data

As the data has been converted to the format of NetCDF, the MATLAB® functions relevant with NetCDF data processing can be adopted easily to extract the data and save the data as
‘mat’ files. Then interpolate the data from the NAE model domain into the POLCOMS model domain.

e) Fill data gaps

As a few gaps exist in the output files of NAE model, in order for the data to be conditioned correctly to POLCOMS, the data which is one-day early than that of missing point, having the same time stamp with the missing data, was used to fill the gaps. Based on this, coupling the gap information produced in step a, a MATLAB® script was created to fill gaps. Fig. A2.6 interprets the procedures of manipulating the BADC meteorological data.
Fig. A2. 6 Flow chart summarising the procedures of producing BADC meteorological data
A3: Computation of Nodal Corrections $u$ and $f$

The nodal corrections $u$ and $f$ must be derived from the Orbital Elements ($p$ and $N$) using the appropriate formulae as follows:

**M1B**: 
\[
\begin{align*}
  f.\sin u &= 2.783 \sin 2p + 0.558 \sin (2p - N) + 0.184 \sin N \\
  f.\cos u &= 1 + 2.783 \cos 2p + 0.558 \cos (2p - N) + 0.184 \cos N
\end{align*}
\]

**M1**: 
\[
\begin{align*}
  f.\sin u &= \sin p + 0.2 \sin (p - N) \\
  f.\cos u &= 2 [\cos p + 0.2 \cos (p - N)]
\end{align*}
\]

**M1A**: 
\[
\begin{align*}
  f.\sin u &= -0.3593 \sin 2p - 0.2 \sin N - 0.066 \sin (2p - N) \\
  f.\cos u &= 1 + 0.3593 \cos 2p + 0.2 \cos N + 0.066 \cos (2p - N)
\end{align*}
\]

**gamma 2**: 
\[
\begin{align*}
  f.\sin u &= 0.147 \sin 2(N - p) \\
  f.\cos u &= 1 + 0.147 \cos 2(N - p)
\end{align*}
\]

**alpha 2**: 
\[
\begin{align*}
  f.\sin u &= -0.0446 \sin (p - p') \\
  f.\cos u &= 1 - 0.0446 \cos (p - p')
\end{align*}
\]

**delta 2**: 
\[
\begin{align*}
  f.\sin u &= 0.477 \sin N \\
  f.\cos u &= 1 - 0.477 \cos N
\end{align*}
\]

**xi 2 / eta 2**: 
\[
\begin{align*}
  f.\sin u &= -0.439 \sin N \\
  f.\cos u &= 1 + 0.439 \cos N
\end{align*}
\]

**L2**: 
\[
\begin{align*}
  &
\end{align*}
\]
\[ f \cdot \sin u = -0.2505 \sin 2p - 0.1102 \sin (2p - N) - 0.0156 \sin (2p - 2N) - 0.037 \sin N\]

\[ f \cdot \cos u = 1 - 0.2505 \cos 2p - 0.1102 \cos (2p - N) - 0.0156 \cos (2p - 2N) - 0.037 \cos N\]

From these formulae the values of \( u \) and \( f \) can be derived.

The formulae for the following fundamental constituents are:

\( u \) of Mm = 0

\( f \) of Mm = 1 − 0.1311 \cos N + 0.0538 \cos 2p + 0.0205 \cos (2p - N)

\( u \) of Mf = −23.7 \sin N + 2.7 \sin 2N − 0.4 \sin 3N

\( f \) of Mf = 1.084 + 0.415 \cos N + 0.039 \cos 2N

\( u \) of O1 = 10.80 \sin N − 1.34 \sin 2N + 0.19 \sin 3N

\( f \) of O1 = 1.0176 + 0.1871 \cos N − 0.0147 \cos 2N

\( u \) of K1 = −8.86 \sin N + 0.68 \sin 2N − 0.07 \sin 3N

\( f \) of K1 = 1.0060 + 0.1150 \cos N − 0.0088 \cos 2N + 0.0006 \cos 3N

\( u \) of J1 = −12.94 \sin N + 1.34 \sin 2N − 0.19 \sin 3N

\( f \) of J1 = 1.1029 + 0.1676 \cos N − 0.0170 \cos 2N + 0.0016 \cos 3N

\( u \) of M2 = −2.14 \sin N

\( f \) of M2 = 1.0007 − 0.0373 \cos N + 0.0002 \cos 2N

\( u \) of K2 = −17.74 \sin N + 0.68 \sin 2N − 0.04 \sin 3N

\( f \) of K2 = 1.0246 + 0.2863 \cos N + 0.0083 \cos 2N − 0.0015 \cos 3N
\[ u \text{ of } M3 = -3.21 \sin N \]

\[ f \text{ of } M3 = (\sqrt{1 + M2})^3 \]

Values for all other constituents can either be derived from the above using the methods described in the next section or else they have values of \( u = 0 \) and \( f = 1 \).

**Derivation of Speeds and values of \( u \) and \( f \) from Constituent Names:**

As shown above the values of \( u \) and \( f \) have been derived from the Orbital Elements for the constituents given, but the values for other constituents can be derived from the construction of the individual constituent names using the principles below.

**Speeds:**

Starting from the left add all the values of the letters of the same name. Therefore, for example, \( MS4 = M2 + S2 \).

But if such addition produces the wrong number of cycles per day, then the signs of the compound constituents must be changed progressively from the right until the correct number of cycles is reached. Thus:

\[ 2MN6 = 2 \times M2 + N2 \] resulting in the correct 6 cycles per day

However, \( 4MN6 = 4 \times M2 + N2 \) which gives incorrect 10 cycles per day, changing the sign of \( N2 \) produces:

\[ 4MN6 = 4 \times M2 - N2 \] which gives the correct value of 6 cycles per day

Some other examples are:

\[ MP1 = M2 - P1 \]
3M2S2 = 3 \times M2 - 2 \times S2

Value of u:

When using the above principles it needs to be borne in mind that u sometimes has a value of zero.

For example, \( u \) of 3M2S2 = 3 \times (u of M2) - 2 \times (u of S2) but u of S2 = 0

therefore, \( u \) of 3M2S2 = 3 \times (u of M2)

Likewise, \( u \) of MP1 = (u of M2) - (u of P1) but u of P1 = 0

therefore, \( u \) of MP1 = u of M2

In addition, because several astronomical constituents have the same values of \( u \) the expression may sometimes be simplified. For example, M2 and N2 have the same value for \( u \) and therefore,

\( u \) of 2MN6 = 3 \times (u of M2)

Value of f:

The values of f are obtained in basically the same manner but multiplying instead of adding the individual contributions. Furthermore, f is always obtained by multiplication and not by division even if a change of sign becomes necessary as explained above.

As with the values of \( u \) the expression if often simplified by the fact that some astronomical constituents have values of \( f = 1 \), and several have the same value.

For example,

\( f \) of MS4 = \( f \) of M2 \times \( f \) of S2 but \( f \) of S2 = 1
therefore, \( f \) of MS4 = \( f \) of M2

\[ f \text{ of } 2\text{MN6} = (f \text{ of } M2)^2 \times f \text{ of } N2 \quad \text{but} \quad f \text{ of } N2 = f \text{ of } M2 \]

therefore, \( f \) of 2MN6 = \( f \) of M2$^3$

\[ f \text{ of } 4\text{MN6} = (f \text{ of } M2)^4 \times f \text{ of } N2 \quad \text{but} \quad f \text{ of } N2 = f \text{ of } M2 \]

therefore, \( f \) of 4MN6 = \( f \) of M2$^5$

\[ f \text{ of } MP1 = f \text{ of } M2 \times f \text{ of } P1 \quad \text{but} \quad f \text{ of } P1 = 1 \]

therefore, \( f \) of MP1 = \( f \) of M2

\[ f \text{ of } 3\text{M2S2} = (f \text{ of } M2)^3 \times (f \text{ of } S2)^2 \quad \text{but} \quad f \text{ of } S2 = 1 \]

therefore, \( f \) of 3M2S2 = \( f \) of M2$^3$

Exceptions:

There are several exceptions to all the above principles. The primary ones are:

MSf:

This has a speed equal to \((S2 - M2)\) and should be treated, therefore, as if it were SMf, and hence the value of \( u \) becomes \((-u \text{ of } M2)\). This will of course have no effect on the value of \( f \), which is always obtained by multiplication and thus equals \( f \text{ of } M2 \).

MA2 and MB2:

Despite their appearance neither of them, nor constituents of other species which include A and B, are compound constituents – there are no constituents A or B to form a compound.
They are constituents whose speeds differ by one cycle a year from that of M2. The A in MA2 was intended to signify the annual differences.

MB2 was originally called Ma2 but this became ambiguous when spoken, or typed on computers without lower case, and so it was initially changed to MA2. However, this in turn was thought to be clumsy and hence MB2 was finally adopted. Although theoretically they should have the same values of $u$ and $f$ as M2, they are so small that they are commonly treated as having values of $u = 0$ and $f = 1$. 
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