An overview of ocean currents with emphasis on currents on the Norwegian continental shelf

Gerhard Ersdal

NPD
March 2001
(preliminary version)

INTRODUCTION

This paper is a summary of a literature study on ocean currents, how currents are generated, what types of currents to expect on the Norwegian continental shelf and the velocities of these currents.

The main sources for this study have been Pond and Pickard (2000), Bearman (1989), Garrison (1999) and Cooper (1986) for general oceanographic information. Sulebak (1991),
Førland (1985) and "Den Norske Los" (1997) are referenced for information on currents on the Norwegian continental shelf.

The current flow varies both in time, with depth below the mean surface and in space. The current is like the wind, being stochastic both in space and time. Current variation, however, stands out from wind variation (and wave variation) as the extreme current velocity is often caused by other mechanisms than what is dominating the current from day to day. The consequence of this is that the extreme values for current may not be predictable from the measurements as the extrapolation to current velocities with small probabilities (long return periods) are dubiously, if the length of the measurements are not sufficient (Myrhaug 1994). The existence of eddies and whirls, described further in the following text, is an example of this.

About 10% of the water in the world are involved in surface currents. Thermal expansion, wind friction and tides drive horizontal flowing water in the uppermost 400 meters of the ocean's surface. Most surface currents move water above the pycnocline. Water within and below the pycnocline (accounting for the remaining 90%) also circulates, but the power for this deeper circulation comes from the action of gravity on adjacent water masses of different densities. Since density is largely a function of temperature and salinity, circulation due to density differences is called thermohaline circulation (Garrison 1999).

**GENERAL OCEAN CIRCULATION**

The water in the oceans is on an everlasting journey driven by the solar energy (giving rise to wind and thermal variations) and tidal forces. These ocean currents are important for the life on the earth. The ocean currents contribute in distributing the energy on the earth, and hence influence heavily on the climate. The ocean currents brings different water masses in contact with each other, and in these contact zones, whirls and eddies will occur that takes the water masses from the surface and down into large depths, which increases the vertical convection.

The ocean currents may be divided into two major circulation systems, surface current circulation and deep-water circulation that are more diffuse and mainly north south directed currents, but these two systems are coupled in many ways.

These two circulation systems will be further described in this section.

---

1 for definitions see vocabulary
In sub-tropic areas of the Atlantic Ocean, Pacific Ocean and in the southern part of the Indian Ocean there are permanent gyres. The gyres are cyclonic (counter-clockwise) in the Southern Hemisphere and anticyclonic (clockwise) in the Northern Hemisphere. The gyres in the Southern and Northern Hemisphere are separated by an eastward current. In all the gyres the western boundary currents are characterized as fast, intense, deep and narrow, while the eastern boundary currents are characterized as slow, wide, shallow and diffuse. The northern gyres is transporting more water and have higher velocities compared to the southern gyres (Sulebak 1991).

In the sub-polar areas on the Northern Hemisphere there are many cyclonic current gyres, where water masses are mixed and homogeneous water are developed. Such sub-polar gyres are not found in the Southern Hemisphere. This is probably due to that the easterly current at 60° S is not forced by continents. The result is a wide circumpolar current around the Antarctica.

Due to the Coriolis force, the currents on the Northern Hemisphere are forced towards the coast, and the coast will always be on the right-hand side of the current. This leads to milder climate on the western parts of continents and colder on the Greenland and Labrador coast.
Figure 2: Global surface circulation in January. Inserted figure shows the currents in northern parts of Indian Ocean during the summer. The numbers are indications of currents referenced in the text (Sulebak 1991).

1: East Greenland Current
The East Greenland Current that is passing along the East Coast of Greenland is a cold return current from the Arctic Ocean. The cold water is passing through the Fram Strait.

2: Labrador Current
The Labrador Current is also a return current bringing cold water from the Arctic along the Labrador Coast and down to the East Cost of Canada and USA.

3: North Atlantic Current
The North Atlantic Current (NAC) is one of the branches of the Gulf Stream. By the time the Gulf Stream reaches the Grand Bank off Now Foundland the Gulf Stream broadened considerable and becomes more diffuse. Beyond this area the Gulf Stream is more correctly referred to as the North Atlantic Current (NAC). Much of the water in the NAC turns southeastwards to contribute to the Canary Current and to circulate again in the subtropical gyre, other flows continue northeastwards into the North Atlantic Ocean. The North Atlantic Current is described further in a later section.

4: Gulf Stream.
The Gulf Stream is in the order of 100 km wide, and has a maximum surface velocity in the order of 3 m/s (Sulebak 1991, p 49). Scour on the seabed under the Gulf Stream shows evidence of a deep current. The volume transport in the Gulf Stream is about 30 Sv in the Stream off Florida, 85 Sv by the time it reaches the latitude of Cape Hatteras. The maximum transport of about 150 Sv is reached at about 65° W (Bearman 1989, p 98). Mesoscale eddies (rings) are formed in the Gulf Stream due to the strong density contrast and velocity shear with respect to surrounding water masses. Warm core eddies may be formed into the American Continental shelf, and cold core eddies is formed into the Sargasso Sea. Cold core eddies may extend to the sea floor at depths of 4000-5000 m, warm core eddies are somewhat shallower but are deep enough to impinge on the continental slope. Cold core eddies typically has a diameter of 150 - 300 km, a warm core eddy has typically a diameter of 100 - 200 km. Velocity in the rings may be as much as 15 - 2 m/s (Bearman 1989).
5: Antilles Current
The Antilles Current is an extension of the Northern Equatorial Current in the Atlantic Ocean. This current is passing east of the Antilles.

6: Canaries Current
The Canaries Current is the diffuse, wide and slow eastern counterpart to the Gulf Stream.

7: Northern Equatorial Current
The Northern Equatorial Current is driven by the trade winds in this area. It is the Southern part of the circulation ring in the North Atlantic Ocean. The Canaries Current, the Gulf Stream and the Northern Equatorial Current is representing the circulation pattern in the Atlantic Ocean.

8: Guyana Current / Northern branch of Southern Equatorial Current
The Guyana Current is a branch of the Southern Equatorial Current, and is coupling with the Northern Equatorial Current north of South America.

10: Southern Equatorial Current
The Southern Equatorial Current (SEC) is, as the Northern Equatorial Current (NEC), driven by the trade winds. The SEC is situated between 20° S and approximately a few degrees north of Equator. The SEC is stronger than the NEC (Sulebak 1991, p 50).

11: Brazil Current
The Southern Equatorial Current is split into two branches outside the coast of South America. The Northern branch is the Guyana Current (8). The Southern Branch is the Brazil Current. The transport in the Brazil Current is in the order of 10-20 Sv, and the current is approximately 100 km wide (Sulebak 1991, p 50).

12: Antarctic Circumpolar Current / West Wind Drift
The Antarctic Circumpolar Current is driven by the dominating westerlies (winds from west in the Ferrel cells) in this area.

13: Benguela Current
The Benguela Current is a branch of the Antarctic Circumpolar Current, forced north along the southwest coast of Africa. The transport in the Benguela Current is in the order of 15 Sv, and the current is more than 500 km wide (Sulebak 1991, p 50).

14: Northeast Monsoon Current
The circulation in the Northern Indian Ocean is different from the general circulation pattern, as the monsoon winds in the summer and winter are dominating the current picture. During the winter, the monsoon winds will be off-shore winds, resulting in the Northeast Monsoon Current.

15: South West Monsoon Current
During the summer, the monsoon winds will be onshore wind, resulting in the South West Monsoon Current.

16: Equatorial Counter Current.
In the Atlantic, Pacific and Indian Ocean there are two easterly currents, counter currents, between the westerly Northern and Southern Equatorial Currents.
- In the Atlantic and Pacific the Equatorial Counter Current is 300-500 km wide. Maximum surface velocity is in the order of 50 cm/s. During the winter (Northern Hemisphere) the current is close to the Equator, approximately at 2° N, and further north during the summer, at approximately 10° N (Sulebak 1991). In the Indian Ocean the Equatorial Counter Current occurs only during the winter (Northern Hemisphere), and is situated south of Equator.
- The Equatorial Under Current is situated at Equator. The center of the current is approximately 100m under the surface. The current is in the order of 30 km wide, but only 200 m deep. The velocity in core may reach 1 - 1.5 m/s (Bearman 1989, p 126). The velocity of the Equatorial Under Current can not be registered at the surface (Sulebak 1991). The Equatorial Under Current is not equally well developed in the Indian Ocean, compared to the Pacific and the Atlantic Ocean.

17: Southern Equatorial Current in the Pacific Ocean
The Southern Equatorial Current (SEC) is, as the Northern Equatorial Current (NEC), driven by the trade winds from east to west. With intervals of 3 to 8 years this trade wind weakens or reverses, giving rise to the El Nino phenomenon.

18: Somali Current
The Somali Current is an analogue to the Gulf Stream in the Indian Ocean. This current only exists during the summer, as the monsoon is an onshore wind. This current is reversed during the winter monsoon.

19: Agulhas Current
The Agulhas Current is the southern branch of the Southern Equatorial Current in the Indian Ocean. This current is running southward east of Africa and west of Madagascar. The current is mixed with the Antarctic Circumpolar Current at approximately 40° S. This current has made this region infamous for giant waves, as waves travelling north-east from the southern Atlantic Ocean tends to be refracted and focused by this current. The volume transport of the Agulhas current is in the order of 130 Sv (Bearman 1989), and is the major western boundary current in the Southern Hemisphere. Eddies like the Gulf Stream rings have also been observed forming in the Agulhas Current. These eddies may be an important agent in the transfer of water between the Indian Ocean and Atlantic Ocean (Bearman 1989).

21: Northern Pacific Current
The Northern Pacific Current is one part of the weak, slow and diffuse eastern boundary part of the northern circulation in the Pacific Ocean.

22: Alaska Current
West of Canada, a branch of the Northern Pacific Current is directed northward, and into the Alaska Bay. This is called the Alaska Current.

23: Aleut Current
The Alaska Current is split into currents at the Aleut, one going further west, and the Aleut Current is going into the Bering Ocean.

24: Oyashio (Oyasiwo)
The Oyashio is moving along the coast of Kamtjatka as a cold return current from the Arctic, as the Labrador Current.
25: Kuroshio (Kurosiwo).
The Kuroshio is, like the Gulf Stream, a narrow and fast current. The maximum surface velocity is in the order of 2 m/s, and the water transport is in the order of 65 Sv (Sulebak 1991, p 50).

26: Northern Equatorial Current in the Pacific Ocean
The Northern Equatorial Current in the Pacific Ocean is situated at the same latitude as the Northern Equatorial Current in the Atlantic Ocean. The current is directed from east towards west, and the transport volume is increasing as it moves west. The current is wide, and the speed seldom exceeds 0.2 m/s, with a transport volume of approximately 45 Sv (Sulebak 1991, p 49).

29: California Current
The California Current and the Northern Pacific Current are the weak, diffuse and broad eastern part of the sub-polar gyre in the Pacific Ocean. The current is approximately 500 - 1000 km wide, and the volume transport is in the order of 10-15 Sv (Sulebak 1991, p 50).

30: Eastern Australian Current
The Eastern Australian Current is a branch of the Southern Equatorial Current in the Pacific Ocean. It passes south along the coast of Australia. The current is approximately 100 km wide, and the volume transport is in the order of 10-25 Sv (Sulebak 1991, p 50). When it reaches New Zealand the current turns towards east and becomes a part of the Antarctic Circumpolar Current.

31: Humbolt Current
The Humbolt Current is, as the Benguela Current, a branch of the Antarctic Circumpolar Current, forced north along the southwest coast of South America and up the coast of Peru and Chile. The transport in the Humbolt Current is in the order of 15-20 Sv, and the current is more than 1000 km wide (Suelbak 1991, p 50).

Deep water (thermohaline) circulation
The surface currents affect the uppermost layer of the worlds ocean (about 10% of its volume), but horizontal and vertical currents also exist below the pycnocline in the oceans deeper waters. The circulation of water at great depths is driven by density differences (convection). Because density is largely a function of water temperature and salinity, the movement of water due to differences in density is called thermohaline circulation (therme = heat, halos = salt).

A general circulation of deep water is generated by the difference in solar heating at the equator and in the polar areas.

According to Garrison (1999) there are five common water masses in temperate and tropical latitudes:
- Surface water, to a depth of about 200 meters
- Central water, to the bottom of the main thermocline (which varies with latitude)
- Intermediate water, to about 1 500 meters
- Deep water, water below intermediate water but not in contact with the bottom, to a depth of 4 000 meters
- Bottom water, water in contact with the seafloor.
The characteristics of each water mass are usually determined by conditions of heating, cooling, evaporation and dilution that occurred at the ocean surface when the mass was formed (Garrison 1999).

During the winter the cooling of surface water in Arctic and Antarctic area, will give relatively dense water that will sink into deep-water masses. Deep water is generally formed in the area between Norway and Greenland (North Atlantic Deep Water), and the Weddel Sea (Antarctic Bottom Water).

The densest and deepest masses were formed by surface conditions that caused water to become very cold and salty. The densest water is formed near the Antarctic coast, and is characterized by a salinity of 34.65 ‰, and a temperature of –0.5°C (Garrison, 1999). The mixture settles along the edge of Antarctica’s continental shelf, descends along the slope, and spreads along the deep-sea bed, moving northward in slow sheets. Antarctic Bottom Water flows many times slower than the water in surface currents. It may take 1 000 years to reach the equator (Garrison 1999).

In the Antarctic convergence zone, downstreaming of water forms the Antarctic Intermediate Water. This water is less dense (lower salinity) compared to the ABW, and sinks to approximately 1000 m.

Some dense bottom water also forms in the northern polar ocean (and Norwegian Sea), but the topography of the Arctic ocean basin prevents most of it from escaping, except in the deep channels formed in the submarine ridges separating Scotland and Iceland and Greenland. These channels allow the cold, dense water formed in the Arctic Ocean and Norwegian Sea to flow into the North Atlantic, forming North Atlantic Deep Water.

In central areas of the northern Norwegian Sea, in between Jan-Mayen and Svalbard, the mixing of arctic water and Atlantic water leads to a water mass of rather homogeneous salinity of 34.9 ‰. When this water is cooled during the winter, this mixed water can obtain such a high density, that it will sink to the bottom. This is occurring in large current whirls in the Greenland Sea and the Norwegian Sea northeast for Island (Gjevik 1996).

A description of the deep water circulation is given by Strommel (Pond and Pickard 2000) and presented in Figure 4.
Figure 4: The deep water is according to Strommels theory transported to the Equator (Pond and Pickard 2000).

The general theory of Strommel is inaccurate, at least in the North Atlantic, and an improved picture of the deep water circulation in this area is given by Hansen and Østerhus (2000). This improved picture is given in Figure 5. This figure shows Arctic Water coming through the Fram Strait and into the North Atlantic, and production areas for overflow water.
Figure 5: Main features of overflow water sources and paths towards the Greenland-Scotland Ridge and paths of the overflow through the eastern North Atlantic and the Charlie Gibbs Fracture Zone (CGFZ). The thickness of the arrows crossing the ridges indicates magnitude and persistence of the overflow (Hansen and Østerhus 2000).
WHAT IS DRIVING THE OCEAN CURRENTS?

Forces acting on the ocean masses

The important forces acting on the water masses can be divided into two groups (Pond and Pickard, 2000), primary forces, which cause motion, and secondary forces, which result from motion. An overview of the forces in the two groups is as follows (Pond and Pickard, 2000, Arntsen and McClimans, 1991 and Myrhaug, 1994):

- **Primary forces**
  - Wind stress (friction from wind)
  - Forces due to variations in surface elevation
    - Gravitation (gravitation from the earth, terrestrial)
    - Tidal (gravitation due to the sun and moon)
    - Atmospheric pressure
    - Water levels changed at river mouths
    - Build up of sea-level along boundaries (coasts)
  - Forces due to variation in density. Density may be changed by:
    - Evaporation - when sea-water evaporates it is leaving a surplus of salt
    - Heating
    - Frost / formation of ice
    - Mixing
    - Melting of ice and snow
    - Influx of water of different density
  - Seismic
  - Slides

- **Secondary forces**
  - Coriolis force (the rotation of the earth)
  - Friction (boundary friction opposing the motion)
  - Topography
  - Stratification

The forces are working on the water masses in the oceans, and forcing them to flow in different patterns. The equation of motion in oceanography (Arntsen and McClimans, 1991):

\[
\frac{\partial \vec{v}}{\partial t} + \vec{v} \nabla \vec{v} + f \times \vec{v} = -\frac{1}{\rho} \nabla p + \nabla (\phi_s + \phi_t) + \vec{F}
\]

\(\vec{v}\) is the velocity vector (x, y and z component of velocity)

\(f\) is the Coriolis coefficient \(f = 2\Omega \sin \phi\), \(\Omega=7.29\times10^{-5}\) rad/s, \(\phi\) is geographic latitude

\(\vec{k}\) is a unit vector in vertical direction

\(\vec{p}\) is pressure vector (x, y and z component of pressure)

\(\phi\) is a potential-vector (gravitational potential from the earth and from tidal forces)

\(\vec{F}\) is a force vector of friction force components from wind on surface, and from bottom and coastal boundaries.
\[
\begin{bmatrix}
  u \\
  v \\
  w \\
\end{bmatrix}
= \begin{bmatrix}
  \frac{\partial}{\partial x} \\
  \frac{\partial}{\partial y} \\
  \frac{\partial}{\partial z} \\
\end{bmatrix}
\n\n\begin{bmatrix}
  0 \\
  1 \\
  -z \cdot g \\
\end{bmatrix}
\begin{bmatrix}
  f_x \\
  f_y \\
  0 \\
\end{bmatrix}
\]

The motions are calculated from the equation of motion, the boundary conditions and the equation of Continuity:

\[
\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0
\]

**Classification of motion**

As an alternative to classify the currents by the forces that act on the water masses, a classification based upon type of motion may be used (Pond and Pickard, 2000) (Myrhaug, 1994) (Bearman, 1989):

- **Tidal motion** (horizontal and internal waves of tidal period)
- **Geostrophic motion** (due to the horizontal pressure gradient force, balanced by the Coriolis force):
  - Inertial motion: motion generated by i.e. a wind blowing steadily in one direction for a time, causing the water to acquire a speed, and then the wind stops and the motion continues without friction as a consequence of its inertia (momentum). Due to the Coriolis force, inertial motion will be in circles: inertial circles.
  - Barotropic motion: Driven by the slope of the water surface causing a horizontal pressure gradient. A barotropic condition is recognized by isopycnic surfaces and isobaric surfaces are parallel and their slope remain constant with depth. The barotropic currents are not varying with depth. Due to the Coriolis force the barotropic current is forced to flow perpendicular to the pressure gradient.
  - Baroclinic motion: Driven by vertical variations in the density of the water (isopycnics). Pressure and density levels are not coincident. A baroclinic conditions is recognized by isopycnic surfaces intersects or are inclined to isobaric surfaces. The baroclinic current is varying with depth. Due to the Coriolis force the baroclinic current is forced to flow perpendicular to the pressure gradient. This includes thermohaline (deep water) circulation.
- **Wind-driven motion** (typical 1-3% of the 1 hour wind speed 10 m above sea level)
- **Internal waves**: Waves, like surface waves that occur at the interface between air and water, can also occur at the boundary between water layers of different densities. These waves are called internal waves.
- **Seismic sea waves**: Waves generated by seismic activity.
- **Various**
  - Gravitational-gyroscopic waves: surface and internal waves of sufficiently long period that the Coriolis effect is important. Wind stress changes and atmospheric pressure changes can cause these waves. Wave periods
approaching one-half pendulum day (Pond and Pickard, 2000, page 207 and 246).

- Kelvin waves: Kelvin waves are generated by a tidal wave or a storm surge. These waves may travel along coasts (with the coast to the right if the direction of travel in the Northern Hemisphere and to the left in the Southern Hemisphere), or they may travel eastward along the Equator as a double wave. In these cases, the coast and the Equator, respectively, are acting as wave guides (Bearman, 1989).

- Rossby waves: Rossby waves result from the need for potential vorticity (rotational momentum) to be conserved. They occur in zonal currents and, relative to the flow, only travel westwards (Bearman, 1989).

- Turbulence: Turbulent motion resulting from velocity shear (change of velocity with respect to one or more spatial coordinates), usually at the water boundaries (Pond and Pickard, 2000, page 14).

The tidal motion is rather easy to forecast, and is regarded as deterministic. The remaining motion components (non-tidal) are called the residual components. These are more difficult to forecast and are often irregular (Førland 1985).

In the following text, the tidal, geostrophic, wind-driven and internal wave motion will be discussed. The remaining motion components will not be covered, although they can be of some importance in some areas. These motion components are further described in Bearman (1989) and Pond and Pickard (2000).

**Tidal motion**

Gravitation due to the moon and the sun, as they rotate about one other in approximately the earth's equatorial plane, is resulting in periodical changes in the water level. The horizontal motion of water due to these water level changes is called tidal motion. Tides may be described as a long wave moving over the oceans. The velocity of these waves are in general depending on the depth, but may reach several hundred meters per second with a wavelength in the order of 5000 nautical miles (Den Norske Los, 1997). The velocity of a tidal wave is slowed down by shallow water.

In a tidal wave the water particles (current) will move in the direction of the wave at the wave crest, and in the opposite direction of the wave in the trough. The maximum current velocities will occur during high and low tides.

A model for understanding amphidromic point is that they are formed when a tidal wave is entering a closed basin. As the wave is entering a the basin (flood tide), the Coriolis force will, in the Northern Hemisphere, direct the wave towards right (from the direction of the water particle). This will result in water piling up at the right boundary (eastern side). When the wave is leaving the basin (ebb tide), the Coriolis force will direct the water towards the opposite boundary (western side). The result is a wave rotating about the amphidromic point.

The oceans of the world are in large and small scales closed basins, and amphidromic points are formed all over the oceans. With a few exceptions, a tidal wave in the Northern Hemisphere will move counterclockwise about amphidromic points and in the Southern Hemisphere the wave will move clockwise about the amphidromic points. Figure 6 shows an
overview of amphidromic points in a course model of the world oceans, and Figure 7 shows more details around the British Isles.

Figure 6: Computer-generated diagram of world-wide amphidromic systems for the dominant semi-diurnal lunar tidal component M2. Blue lines are co-range lines and red lines are co-tidal lines (Bearman 1999).

The paths of rotation of the sun and the moon about the earth are not circles but ellipses, and the planes of rotation are not always in equatorial plane but move north and south with an annual cycle for the sun and a monthly cycle for the moon. These facts add further complications to the resultant tide-producing forces. In addition, longer period variation up to 19 years period occurs due to other motions (Pond and Pickard, 2000, p 259). These constituents are usually divided into three groups, semi-diurnal (one-half day), diurnal (one day) and longer period. A harmonic model is usually used to represent the tidal waves. The total tidal water is a result of the sum of several individual waves, resulting from the different motions of the moon and the sun. The total tidal water level is represented by the equation:

\[ h(t) = MV + \sum_j f_j \cdot H_j \cdot \cos(\sigma_j \cdot t + (V_0 + u)_j - g_j) \]

- \( h(t) \): water level
- \( MV \): mean water level
- \( f_j \): correction factor for the 18.6 yearly variation for component j
- \( H_j \): amplitude of component j
- \( \sigma_j \): frequency for component j
- \( t \): time
- \( (V_0+u)_j \): phase at t=0
- \( g_j \): phaseshift of component j
Figure 7: Amphidromic systems around the British Isles. The figures on the co-tidal lines (red) indicate the time of high water (in hours) after the moon has passed the Greenwich meridian. Blue lines are co-range, with tidal range in meters (Bearman 1999).
The most important components of the tidal water level are represented in Table 1.

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Period (hours)</th>
<th>Coefficient ratio</th>
</tr>
</thead>
<tbody>
<tr>
<td>Semi-diurnal (twice daily):</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Principal lunar</td>
<td>M₂</td>
<td>12.42</td>
</tr>
<tr>
<td>Principal solar</td>
<td>S₂</td>
<td>12.00</td>
</tr>
<tr>
<td>Larger lunar elliptic</td>
<td>N₂</td>
<td>12.66</td>
</tr>
<tr>
<td>Luni-solar</td>
<td>K₂</td>
<td>11.97</td>
</tr>
<tr>
<td>Diurnal (daily):</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Luni-solar</td>
<td>K₁</td>
<td>23.93</td>
</tr>
<tr>
<td>Principal lunar</td>
<td>O₁</td>
<td>25.82</td>
</tr>
<tr>
<td>Principal solar</td>
<td>P₁</td>
<td>24.07</td>
</tr>
<tr>
<td>Longer period:</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lunar fortnightly</td>
<td>Mₕ</td>
<td>327.86</td>
</tr>
<tr>
<td>Lunar monthly</td>
<td>Mₘ</td>
<td>661.3</td>
</tr>
</tbody>
</table>

Measurements of tidal variations should extend the 18.6 years variation, and last for at least 19 years, or be corrected for this variation.

Tidal currents are regular and predictable and the maximum tidal current is associated with the highest or lowest astronomical tide, HAT or LAT. They are generally weak in deep water past the shelf break, in the order of 0.1-0.3 m/s. Tidal currents can be strengthened by shoreline or bottom configurations such that strong tidal currents can exist in many inlets and coastal regions. A maximum of 8 m/s for Seymour Narrows in western Canada is frequently quoted as one of the highest speeds measured (Pond and Pickard 2000, p 277).

**Geostrophic motion**

In the same way as wind blows from high to low pressure (but is restricted by the Coriolis force to circulate instead of blowing directly from the high pressure to the low pressure), water tends to flow to even out lateral differences in pressure. The force that gives rise to this motion is known as the horizontal pressure gradient force. If the Coriolis force acting on moving water is balanced by a horizontal pressure gradient force, the current is said to be in geostrophic equilibrium and is described as a geostrophic current (Bearman 1989). Figure 8 is illustrating the forces acting on the water masses resulting in geostrophic motion.
Figure 8: In the Northern Hemisphere, a sea-surface upward slope towards the east results in a horizontal pressure gradient toward the west. Initially this causes motion "down the pressure gradient", but because the Coriolis force act at right angles to the direction of the motion, the equilibrium situation is one in which the direction of the flow is at right angles to the pressure gradient (Bearman, 1989).

Pressure differences are created in the oceans when the sea-surface is higher at one point compared to another. In this case the surface of equal pressure within the ocean - isobaric surfaces - are parallel to the sea-surface. If the ocean water is well-mixed and therefore fairly homogeneous, the density will increase with depth because of the compression caused by the weight of overlaying water. As a result, the isobaric surfaces are parallel not only to the sea-surface, but also to the surfaces of constant density or isopycnic surfaces. Such conditions are described as barotropic (see Figure 9), giving rise to what is called a barotropic current, which is constant over the water depth.
Figure 9: In barotropic flow, isopycnic surfaces and isobaric surfaces are parallel and their slopes remain constant with depth. Because the slope of the isobaric surfaces remains constant with depth, the horizontal pressure gradients from B to A, and hence the geostrophic current, is constant with depth (Bearman, 1989).

Barotropic conditions leading to geostrophic currents are due to variations in surface elevation, in general caused by:
- Variation in atmospheric pressure
- Water levels changed at river mouths
- Build up of sea-level along boundaries (coasts) due to wind stress

In most situations, there are lateral variations in density, and isobaric surfaces are not parallel to the sea-surface. Isobaric surfaces intersect isopycnic surfaces and the two slopes in opposite directions. Because isobaric surfaces are inclined with respect to one another, such conditions are known as **baroclinic** (see Figure 10), giving rise to a **baroclinic current**, which is varying with depth.
Figure 10: In baroclinic flow, the isopycnic surfaces intersect (or are inclined to) isobaric surfaces. At shallow depths, isobaric surfaces are parallel to the sea-surface, but with increasing depth their slope become smaller, because the average density of a column of water at A is more than a column of water at B. As the isobaric surfaces become increasingly near horizontal, so the horizontal pressure gradients decreases and so does the geostrophic current, until at some depth the isobaric surfaces are horizontal and the geostrophic current is zero (Bearman, 1989).

Baroclinic conditions leading to geostrophic currents are due to variation in water density, in general caused by:

- Evaporation - when sea-water evaporates it is leaving a surplus of salt
- Heating
- Frost / formation of ice
- Mixing
- Melting of ice and snow
- Influx of water of different density by large current velocity

Sea-surface slopes associated with geostrophic currents are broad, shallow, topographic irregularities. They may be caused by prevailing winds "piling up" water against coastal boundaries, by variation in pressure in the overlying atmosphere, or by lateral variations in water density resulting from differing temperature and salinity characteristics, or by a combination of these. The slope gradients of about 1 to $10^5$ to 1 to $10^8$, i.e. a few meters in $10^2$ - $10^5$ km, so they are extremely difficult to detect and measure (Bearman 1989). However, under baroclinic conditions the isopycnic surfaces may have slopes several hundred times this, and these can be measured (Bearman 1989).

The density of water is difficult to measure directly. To cope with this problem, temperature and salinity are measured - and the density is calculated based on these measurements. If the
distribution of density is known over a section, the slope of the velocity can be calculated, but the absolute values of current velocity must be estimated for one of the levels by different methods (Førland 1985).

Circulation currents are relative steady, large scale features of the general oceanic circulation. While relative steady, these circulation features can meander and intermittently break off from the main circulation feature to become large scale eddies or rings, which then drift at a few miles per day (ISO 19901-1). Such eddies, rings or whirls are also seen in the Norwegian Coastal Current (NCC) and the North Atlantic Current (NAC). These features are discussed in more details in a separate section.

**Wind-driven motion**

The classic theory for wind-induced currents is that of Ekman in 1905, stating that in the Northern hemisphere the wind-induced current is directed 45° to the right of the wind direction at the ocean surface, and rotates further clockwise with increasing depth (Figure 11 illustrates this phenomenon). According to Pond and Pickard (2000), this is a mathematical solution by Ekman to the observations and qualitative theory by the Norwegian explorer Fridtjof Nansen. The surface current magnitude is predicted to be proportional to the wind stress, and inversely proportional to the square root of the eddy viscosity. Since then, different estimates of the wind-induced current have been reported, depending in the formulation of the eddy viscosity, in particular its vertical variation (Lønseth et al, 1988).

![Figure 11: The Ekman spiral current pattern believed to result from the action of wind on the surface. The lengths and directions of the black arrows representing the speed and direction of the black arrows represent the speed and direction of the wind-driven current. For the Ekman layer as a whole, the force due to the wind is balanced by the Coriolis force, which in the Northern Hemisphere is 90° to the right of the wind direction (Bearman, 1989).](image)

Observations of wind-driven current in real oceans, away from coastal boundaries, have shown that the surface currents are similar to those predicted by Ekman. However, the deviation is less than the 45° predicted by the Ekman theory (Bearman, 1989). A deviation in the order of 20 - 30° (increasing with increasing wind speed) seems to be appropriate (Førland
1985). This fits well with the Nansen's original observations and qualitative explanation from 1898 (Pond and Pickard 2000, p 101) stating a angle of 20 - 40° between the wind and current.

Vyas et al (1988) are describing the measurements performed on the Odin platform in the North-Sea during the winters of 1984/85, 1985/86 and 1986/87. They are concluding from these measurements that "The data demonstrates consequently that a uniform, slab profile of steady current is an excellent engineering model for severe storms in well-mixed oceans, such as the North-Sea and Norwegian Sea in winter". The idea behind this assumption is that high wind is causing turbulence in the water and quickly eliminates the density variations causing the current. This is giving a situation where high waves and high geostrophic current can not occur at the same time (Arntsen and Mcclimans, 1991). This statement will only apply for water depths up to approximately 100 m, as water beneath this level will not be influenced by the wind at the surface.

Cooper (1986) describes these two theories as two primary schools of thoughts regarding how the wind stress is transferred. One is claiming that the momentum is transferred through the mixed layer within an hour, resulting in a nearly constant velocity throughout the upper layer (often referred as slab flow). Others argue that the transfer of momentum is much slower, giving rise to vertical gradients of the horizontal velocity components, called the Ekman-flow. These two theories exist primary because the available database is inadequate (Cooper, 1986).

In addition to the current resulting from wind drag, the waves will induce a current, often called Stokes drift (Førland 1985). If linear wave-theory is used for calculating particle velocity, particles will move in closed elliptic movements, and will not result in any current. If higher order theory is used and turbulent friction is included, calculations show that the wave movement is resulting in a mean forward velocity (Førland 1985).

The wind stress at the sea-surface not only causes horizontal movement of the water, it also induces vertical movement in the surface water. When the wind stress leads to a divergence of surface water, deeper water rises up to take its place (upwelling). Convergence of surface water leads to sinking of surface water (downwelling). Upwelling brings deep, cold, usually nutrient-laden water towards the surface. Downwelling helps supply the deeper ocean with dissolved gasses and nutrients and assists in the distribution of living organisms. It should also be mentioned that divergences and convergences might also be a result of fronts between opposing currents. The rings in the Gulf Stream are an example of this.

Upwelling and downwelling occur throughout the oceans, at coastal boundaries, as a result of cyclonic and anticyclonic winds, in the Equatorial Divergences and Convergence's and as a result of Langmuir Circulation. A further description of these vertical movements of the water due to wind stress at sea-surface is given in Bearman (1989).

Internal waves

Waves, like surface waves that occur at the interface between air and water, can also occur at the boundary between water layers of different densities. As is the case with ocean waves at the air-ocean interface, internal waves possess troughs, crests, wavelengths and periods. Internal waves also share other common characteristics with surface wind waves as breaking, non-linear wave interaction and reflection (Cooper, 1986). Surface waves move rapidly because the difference in density between air and ocean is great (1000 times). Internal waves usually move very slowly because the density difference between the nearby media is very
small. Internal waves occur in the ocean at the base of the pycnocline, especially at the bottom edge of a steep thermocline (Garrison, 1999).

According to Martin et al (1994), the first recorded observations of internal waves were made as a result of Fridtjof Nansen's expedition to the Arctic in 1893. Nansen's ship, Fram, experienced increased drag while sailing across what he called "dead water", which consisted of a thin layer of fresh water overlying salty water. The Fram was generating internal waves in the interface between the fresh and salty layers, and these waves were propagating energy away from the ship.

The wave height of internal waves may be greater than 30 meters (Garrison, 1999), with a wavelength that often exceeds 0.8 kilometers (Garrison, 1999) and a period typically in the order of 5-8 minutes (Garrison, 1999) or as high as 10 minutes to 1 day (Cooper, 1986). In ocean waters where density differences are smaller, periods of up to 12 hours and amplitudes of 10 to 300 meters or more have been recorded (Pond and Pickard, 2000). Cooper (1986) indicates that internal waves can generate currents in the order of 0.5 m/s at frequencies as low as 5 cycles per hour.

Internal waves can be generated by a number of mechanisms including atmospheric forcing (i.e. by travelling pressure fields or variable wind stress), surface waves and topography in conjunction with quasi-steady currents (Cooper, 1986). The longer period internal waves, containing significant energies, tend to be forced oscillations driven by the astronomical tide in its interaction with the local topography. This type of wave is often termed an internal tide. According to Garrison (1999), oceanographers are not certain how these large, slow motion waves are generated, but wind energy, tidal energy or ocean currents may be responsible.

Two types of internal waves are recognized, the linear internal wave - a regular wave with many waves following each other, and the solitary internal wave (solitons) - a single crest wave (Martin et al, 1994). Generation of internal solitons appears to be by tidal flow over a submerged ridge, a trapped internal lee wave forms behind the ridge and the energy it contains is transferred into a packet of solitons of varying sizes travelling upstream as the tide slackens (Martin et al, 1994).

The first known implication of internal waves on the offshore industry, occurred in the Andaman Sea (Osborne and Burch, 1980). High velocity currents where generated by internal soliton waves with an amplitude of 60 to 90 m with a period of 40 minutes or less. Each wave train consisted of approximately 5 solitons. The length of each wave varied between 600 to 1200 m. The maximum registered current velocity was 1.8 m/s. The phenomenon was linked to the tidal current, as it occurred at the turn of the tide.

An overview of observations of internal soliton waves is compiled by Roth (2000), and parts of these data are reproduced in Table 2.
Table 2: Observations of internal soliton waves. A: amplitude, U: current velocity, L: length scale, P: period, N: numbers of waves in each wavetrain, C: phase velocity (Roth, 2000)

<table>
<thead>
<tr>
<th>Reference</th>
<th>A (m)</th>
<th>U (m/s)</th>
<th>L (m)</th>
<th>P (min)</th>
<th>N</th>
<th>C (m/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Osborne and Burch (1980)</td>
<td>60</td>
<td>1.8</td>
<td>&lt;15 000</td>
<td>40</td>
<td>5</td>
<td>2.14</td>
</tr>
<tr>
<td>Apel et al (1985)</td>
<td>&lt;90</td>
<td>&lt;1.1</td>
<td>6 300</td>
<td>45</td>
<td>3-10</td>
<td>&lt;2.6</td>
</tr>
<tr>
<td>Holloway (1987)</td>
<td>30</td>
<td>-</td>
<td>428</td>
<td>20</td>
<td>12</td>
<td>&lt;0.92</td>
</tr>
<tr>
<td>Nagovitsyn et al (1991)</td>
<td>5-10</td>
<td>-</td>
<td>-</td>
<td>10-15</td>
<td>-</td>
<td>0.7</td>
</tr>
<tr>
<td>Pingree and Mardell (1985)</td>
<td>30</td>
<td>-</td>
<td>1000</td>
<td>-</td>
<td>-</td>
<td>0.7</td>
</tr>
<tr>
<td>Holligan et al (1985)</td>
<td>50-80</td>
<td>-</td>
<td>1000-1500</td>
<td>-</td>
<td>2-6</td>
<td>1.1</td>
</tr>
<tr>
<td>New and Pingree (1990)</td>
<td>60-70</td>
<td>&gt;0.85</td>
<td>1000-2000</td>
<td>20-30</td>
<td>2-8</td>
<td>1.0</td>
</tr>
<tr>
<td>Allen (1983)</td>
<td>-</td>
<td>-</td>
<td>500-1000</td>
<td>-</td>
<td>4-10</td>
<td>-</td>
</tr>
<tr>
<td>Watson and Robinson (1990)</td>
<td>&lt;60</td>
<td>1</td>
<td>500-1000</td>
<td>10-15</td>
<td>1-7</td>
<td>&lt;2.5</td>
</tr>
<tr>
<td>Sapia and Salutsi (1987)</td>
<td>10-17</td>
<td>-</td>
<td>-</td>
<td>2-7</td>
<td>-</td>
<td>0.72</td>
</tr>
<tr>
<td>Halpern (1971)</td>
<td>&lt;10</td>
<td>-</td>
<td>200</td>
<td>6-8</td>
<td>-</td>
<td>0.88</td>
</tr>
<tr>
<td>Haury et al (1983)</td>
<td>&lt;30</td>
<td>-</td>
<td>300</td>
<td>8-10</td>
<td>-</td>
<td>0.6</td>
</tr>
<tr>
<td>Sandstrom and Elliot (1984)</td>
<td>60</td>
<td>-</td>
<td>-</td>
<td>2-4</td>
<td>1.0</td>
<td></td>
</tr>
<tr>
<td>Sandstrom et al (1989)</td>
<td>30</td>
<td>-</td>
<td>-</td>
<td>7-8</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td>Brickman adn Loder (1993)</td>
<td>20</td>
<td>-</td>
<td>2000</td>
<td>-</td>
<td>-</td>
<td>0.4</td>
</tr>
<tr>
<td>Zheng et al (1994)</td>
<td>5.6</td>
<td>0.13</td>
<td>600</td>
<td>23.8</td>
<td>5</td>
<td>0.42</td>
</tr>
<tr>
<td>Gaporovic et al (1988)</td>
<td>8</td>
<td>0.42</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.7</td>
</tr>
<tr>
<td>Gan and Ingram (1992)</td>
<td>2-3</td>
<td>-</td>
<td>-</td>
<td>2-6</td>
<td>3-6</td>
<td></td>
</tr>
<tr>
<td>Gargett (1976)</td>
<td>-</td>
<td>&lt;0.25</td>
<td>50-100</td>
<td>-</td>
<td>-</td>
<td>0.25</td>
</tr>
<tr>
<td>Howell and Brown (1985)</td>
<td>29</td>
<td>0.14</td>
<td>-</td>
<td>28</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td>Fu and Holt (1984)</td>
<td>50</td>
<td>-</td>
<td>&lt;2000</td>
<td>-</td>
<td>10-20</td>
<td>1.2</td>
</tr>
<tr>
<td>Apel and Gonzalez (1983)</td>
<td>56</td>
<td>-</td>
<td>200-1600</td>
<td>-</td>
<td>2-20</td>
<td>-</td>
</tr>
</tbody>
</table>

According to Garrison (1999) internal waves may have been a participating cause of the loss of the submarine USS Tresher in 1963, 350 km east of Boston (Jackson, 2000), causing the loss of 129 lives.

**Storm surge**

The bulge of water driven by a storm or tropical cyclone is called a storm surge. The low atmospheric pressure associated with great storm will draw the ocean surface into a broad dome as much as 1 meter (Garrison, 1999). According to Arntsen and Mcclimans (1991) a storm can give impulse to a surge that may propagate away from the storm area, or in front of the storm area.

According to Cooper (1986, page 191) the storm surge observed in shallow water is a forced long wave. Such waves can also be important in the shelf slope areas, generating currents in the order of 0.5 m/s.

Barltrop (1998, page 2-41) describes storm surge as follows: "Surge is the name given to changes in water level due to meteorological forcing which is the combined effect of wind and atmospheric pressure. Storm surge is the surge associated with a storm and ‘storm tide’"
has come to mean the combined effects of storm surge and astronomical tide. During a storm event, varying surge levels will occur over large areas and some locations may actually experience negative surges phenomenon which can be important when navigating in shallow water."

In the Northern Hemisphere, a low-pressure field will give cyclonic winds. Due to the Ekman transport (the average movement of the wind-driven layer is 90° to the right), the wind will cause a divergence of surface water and upwelling (of the deep water). Under these conditions, the sea-surface is lowered and the thermocline is raised. This may give rise to the negative storm surge registered in some measurements. The raised thermocline can also be the starting point of an internal wave.

The total picture of the currents due to a storm also needs to take into atmospheric pressure gradients throughout the storm. The total current velocities are a complex function of the storm strength and meteorological characteristics, bathymetry and shoreline configuration, and water density profile. In deep water along open coastlines, surface current can be roughly estimated to have velocities up to 1-3% of the 1 hour sustained wind velocity during storms. As the storm approaches the coastline and shallower water, the storm surge and current can increase.
Why does the global surface circulation patterns form?

The global surface circulation in the oceans is driven by the solar energy, which gives rise to wind and thermal variations, and tidal forces. The general wind patterns are described in Figure 12.

Figure 12: Global air circulation as described in the six-cell circulation model (Garrison 1999).

The surface winds that form global patterns within latitude bands, is presented in Figure 12. Most of Earth's surface wind energy is concentrated in each hemisphere's trade winds, easterlies and westerlies. Waves on the sea surface transfer some of the energy from the moving air to the water by friction. This tug of wind on the ocean surface begins a more rapid mass flow of water. As a rule of thumb, the friction of wind blowing for at least 10 hours will cause surface water to flow downwind at about 1-3% of wind speed.

The moving water will "pile up" in the direction the wind is blowing. Water pressure will be higher in the "piled up" side, and the force of gravity will act to pull the water down the slope - against the pressure gradient - in the direction from which it came. But the Coriolis effect intervenes. Because of the Coriolis effect, Northern Hemisphere surface currents flow to the right of the wind direction. Southern Hemisphere currents flow to the left (Garrison 1999). The major ocean circulations, like the Gulf Stream, are geostrophic currents resulting from the 'pile up' of water caused by the prevailing westerly trade winds. Flow in the Gulf Stream is in approximate geostrophic equilibrium, and the strong lateral gradients in temperature and salinity mean that the flow is baroclinic (Bearman 1989, p 119).
Solar heating causes water to expand slightly. Because of this, the sea level near the equator is about 8 centimetres higher than sea level in temperate ocean areas. Water in the cold regions near the poles cools and contracts by a similar amount. This global difference creates a very slight slope, and warm equatorial water flows "downhill" (polewards) in response to gravity. Poleward moving surface water tends to lag behind the Earth's eastward rotation, however, and moves toward the ocean's western boundaries. The water's slow travel is also influenced by the Coriolis effect. A sluggish circular flow therefore develops in ocean basins on either side of the equator. The difference in temperature between polar and tropical regions is, however, not the major primary force responsible for the surface currents (Garrison 1999).

**Whirls and eddies in the currents**

Circulation currents are relatively steady, large-scale features of the general oceanic circulation. Examples include the Gulf Stream in the Atlantic Ocean and the Loop Current in the Gulf of Mexico, where surface velocity can be in the range of 1 to 2 m/s. While relatively steady, these circulation features can meander and intermittently break off from the main circulation feature to become large scale eddies or rings, which then drift at a speed of a few miles per day. Velocity in such eddies or rings in the Gulf Stream east of USA, can approach those of the main circulation feature. At other locations, i.e. at the Troll field west of Norway, velocities in the order of 1.8 m/s (Lønseth 1988) are measured in the Norwegian Coastal Current. The Norwegian Coastal Current has a mean velocity of approximately 0.4-0.5 m/s (Førland 1985). These circulation features and associated eddies usually occur in deep water beyond the shelf break and generally do not affect sites with depths less than approximately 200 m.

Barltrop (1998, page 2-37) describes this phenomenon as follows: “... baroclinic currents tend to become unstable so that eddies form. Examples of areas where eddies are of importance to the offshore industry are the waters off Norway and in the Gulf of Mexico. Large scale meandering in the Norwegian current can be caused by a sudden outflow of brackish water from the Baltic into the North Sea. This can occur following the end of a period of strong westerly winds and the moving water of different density from that of the North Sea generates eddies with diameters up to 100 km. Current speeds within the eddy exceed 1 m/s and may arrive suddenly without warning.”
CURRENTS ON THE NORWEGIAN CONTINENTAL SHELF

The ocean on the Norwegian continental shelf contains parts of Skagerrak, North Sea, Norwegian Sea, Barents Sea and parts of the Arctic Ocean. The dominating currents and important phenomenon on the Norwegian continental shelf are:

- North Atlantic Current
- Norwegian Coastal Current
- Inflow of Polar water
- Wind driven current
- Tidal current (up to 0.5 m/s according to NORSOK N-003)
- Eddies in the Norwegian coastal current from Lista to Stadt (up to 1.8 m/s according to Lønseth 1988)
- Eddies in the North Atlantic Current (Ormen Lange, Helland Hansen, Svinøy)
- Possible internal waves of the shelf ridge

The currents on the Norwegian Continental Shelf are in general controlled by the bottom topography. A 3D presentation of the topography of North Atlantic is shown in Figure 13.

![Figure 13, Topography of the North Atlantic (Gjevik 1996).](image)

The dominating mean currents on the Norwegian Continental Shelf are described in many references, and details may differ between these references. Figure 14 shows one of these representations, taken from “Den Norske Los”, a guidance given for boat pilots. Hansen and Østhus (2000) give a slightly different picture of the current system, as shown in Figure 15, with some more details on the Atlantic inflow. Further, details on the currents in the North Sea basin are given in Figure 16.
Figure 14: Surface currents on the Norwegian continental shelf (Den Norske los 1997)
Figure 15: Main features of the near-surface circulation in the eastern North Atlantic and the Nordic Seas. Continuous arrows show Atlantic water flow. Broken and dotted arrows indicate flow of other water masses. Water masses transported by the main current branches are indicated (Hansen and Østerhus 2000).
These figures (Figure 14, Figure 15 and Figure 16) clearly show the current pattern dependency on bottom topography.

The main current circulation system is described here in the order as they are numbered in Figure 14, with (heavy) emphasis on the North Atlantic Current and Norwegian Coastal Current.

1: The North Atlantic Current (NAC)

The North Atlantic Current (NAC) is entering the Norwegian Sea through the Shetland-Faeroe-Islands Trench and to a smaller degree, over the ridge between the Faeroe-islands and Island. This second branch of NAC is not shown in Figure 14, but is included in Figure 15. Measurements with drogued drifters (Poulain 1995) also clearly shows the same second branch (see Figure 17).
Figure 17: Trajectory segments corresponding to fast currents (>40 cm/s) as measured by drogued drifters. Bathymetry is depicted with gray shades (Poulain 1995).

The NAC is to be considered as a topographically controlled flow, with the core of the eastern branch following the 500 m contour (maximum speed of 117 cm/s), and the western branch following the 2000 - 2500 m contour (maximum speed of 87 cm/s) at the Svinøy Section (Gjevik 2000, p 13).

The main part of the NAC is following the continental shelf slope northwards along the Norwegian coast towards Storegga. The current is following the continental slope further north. At Tromsøflaket the current is splitting into two parts, where one part is entering the Barents sea and the other part is following the continental slope northwards towards Svalbard and further into the Arctic Ocean. The water transport is varying with seasons, but also on shorter time-scales. Estimates are according to Gjevik (1996) indicating a water flow of 3 - 12 Sv. Typical velocities in this current are 0.2 - 0.4 m/s.
Eddies in the North Atlantic Current (NAC)
Eddies in the NAC are seen both on measurements and on satellite observations (SAR data from the European ERS-2 satellite). Statistics of these eddies are not fully established, but a reasonable new summary is given by Gjevik (2000).

Atlantic Water in the North Sea
Two (three according to Figure 15, but the two most northern are treated as one) branches of the NAC is entering the North Sea basin between northern Scotland and western Norway. These two branches are:
- **The Shetland Current (Atlantic Inflow Current)** is a branch of the Norwegian Atlantic Current and is considered to be a topographically controlled flow which is steered along the slope of the Tampen plateau to the north of the Shetland Islands. Then following the western slope into the Skagerrak, it is becoming gradually fresher and cooler due to lateral mixing. The water flow in this current is estimated to 1.1 Sv. Typical velocities are 0.1-0.2 m/s, but velocities at 0.8 m/s are measured near the bottom in the western slope (Førland 1985). This current is marked as SC (Shetland Current) in Figure 15.
- **The Fair Island Current** is according to Førland (1985) a current is entering the North Sea between the Orkney Islands and Shetland. This current is called the Fair Island Current. The water flow is estimated to 0.3 Sv, but the flow is narrow and resulting in current velocities of 0.25 m/s. This current is marked in Figure 15 as the lowest of the three branches of the Shetland Current.

2: The Norwegian Coastal Current (NCC)
The Norwegian Coastal Current (NCC) is the result of the outflow of the mixture between Baltic water (fresh water from the rivers of Northern Europe) and the North Sea water which has entered the Skagerrak along the west coast of Denmark in the Juntland Current. More fresh water is added to the Skagerrak from rivers. Further mixing increases the volume of brackish water and eventually this water flows out of the Skagerrak and northward along the coast in the NCC. Additional fresh water is added to this current from Norwegian rivers. The NCC follows the Norwegian coast. The salinity increases from about 30 ‰ in the Skagerrak to approximately 34 ‰ at the North Cape. Coastal water is characterized by salinities below 34.7 ‰.

NCC follows the Norwegian coast from the inner parts of Skagerrak, around the south coast of Norway and turns into a northerly direction along the West coast. The velocity of the current is variable and may possibly amount to 0.5 m/s on average for a week or two in favorable weather conditions. Current speeds exceeding 1 m/s are frequently observed.

Whirls in the Norwegian Coastal Current
During the development of the Troll field, west of Bergen, an investigation into the whirls in the NCC was performed. Whirls were used as a term describing clockwise eddy motions. The term eddy was reserved for counter-clockwise eddy motions, which often penetrates to larger depth.

Atmospheric pressure patterns and storm surges cause a time varying outflow. The time-changing outflow from the Skagerrak is the most significant variable for fluctuation in the NCC along the West Coast of Norway up to Stad. Significant mesoscale eddies near the Troll Field are associated with large outflows on the order of 1 Sv. The average outflow from Skagerrak is less than 0.25 Sv (Lønseth, 1988).
The development of these meanders usually starts on the south-west coast, and they deteriorate as they reach the area around Stad, at 62° N. Current speed in excess of 1.5 m/s have been measured in the surface in these whirls.

My understanding of the generating of these whirls are:
- earth rotation and shear between nearby water masses leads to meandering
- the area where the meandering is created is a frontal area with NCC going north and NAC going south west of NCC
- these meanderings lead to whirls

3: The Nordkapp Current
The Nordkapp current flows into the Barents Sea where it divides into several branches. The Atlantic water (in the Nordkapp Current) which flows northwards in Hopenydepetus the Polar Water towards Storbanken and because its greater density sinks down from the surface and continues beneath the Polar Water. The mean current speed at Sentralbanken is in the order of 22 - 24 cm/s and a maximum current of 40 - 50 cm/s at all depths (Bjerke and Torsethaugen 1989).

4: South Spitsbergen Current
A branch of the NAC.

5: The West Spitsbergen Current
West of Spitsbergen the NAC is named The West Spitsbergen Current.

6: The Percy Current
The most stable of the north-easterly cold currents entering the Barents Sea. The current is entering the Barents Sea between Fraz Josefs Land and Novaja Semlja.

7: The Bjørnøya Current
The Bjørnøya current is a rather small and narrow current following the southern slope of the Svalbard Banks. The Bjørnøya current consists of Polar Water which comes from the East Spitsbergen Current and the Percy Current from the area south of Franz Josefs Land.

8: The East Spitsbergen Current
North of the Bjørnøya Current, and parallell to this, the East Spitsbergen Current runs between Hopen and Edge Island. Much of it turns at South Cape, and flows northwards along the west coast as the Sørkapp Current. This current carries polar and Arctic water with temperatures below 0° C and low salinity. This current carries ice along the coast, especially early in the summer (Bjerke and Torsethaugen 1989).

9: The Sørkap Current
See description of the East Spitsbergen Current
Tidal current on the Norwegian continental shelf

The tidal currents on the Norwegian continental shelf are thoroughly described in Gjevik et al (1990). In NORSOK N-003 Figure 18 is given as a general overview of tidal currents.

Figure 18, Maximum 100 year tidal surface current in m/s (NORSOK N-003)
THE EFFECT OF CURRENTS ON OFFSHORE STRUCTURES

The implication of current on offshore structures will vary according to the water depth at the specific site and the type of offshore structure. A few general effects on offshore structures will be mentioned here, but not described in any detail.

- On Jacket structures on the Norwegian Continental shelf current is contributing with 20 - 40 % of the base shear, depending on the size of the jacket and the current speed.
- On flexible risers and pipelines the current is resulting in vortex shedding induces vibrations (VIV).
- Mooring lines are forced by current.
- Ship shaped installations (FPSO, FSU etc.) experience fish-tailing due to current.
- Internal waves set up rather high current velocities, and may influence with drilling risers, flexible riser and mooring lines.
- Currents along the coast and in fjords will influence the design of pipelines.

The effect of current will be very dependent on the depth on the actual site. Shallow water currents are dominated by wind friction, astronomical tides and the two boundaries - the sea floor and the coastline. Deepwater currents are (in contrast to shallow water currents) largely unaffected by bottom friction and tides. The importance of the coastline is replaced by the continental slope, which serves as a guide to focus current energy. Density stratification is of critical importance in deep water. It appears in the form of "fronts" in the horizontal and as a thermocline or pycnocline in the vertical (Cooper 1986).

Grant et al (1995) describe the current at the Foinhaven field in relative deep water north of Scotland, and compare these data with the Magnus field. One of the remarkable current phenomenon at the Foinheaven field is the reversed (soutward) deep water current of 1.15 m/s (extrapolated 100 years value) 4 meters above the seabed. The surface current is dominated by the North Atlantic Drift, with a maximum value of 2 m/s (extrapolated 100 years value).

NORSOK N-003 recommends evaluating the following current components for structural design:
   a) wind induced current
   b) tidal current
   c) coast and ocean currents
   d) local eddy currents
   e) currents over steep slopes
   f) currents caused by storm surge
   g) internal waves

It is also appropriate to quote ISO/WD 19901-1. "The characteristics of the extreme current profile that need to be estimated for the design of offshore structures are particularly difficult to determine since current measurement surveys are relatively expensive and consequently it is unlikely that any measurement program will be sufficiently long to capture a severe event. Furthermore, current (hindcast) modeling is not as advanced as wind and wave modeling in terms of being able to provide the parameters needed."
However, the development of hindcast models for current has improved rather significant over the last years.

**ACKNOWLEDGEMENT**

I would like to thank Arne Kvitrud, Ove Tobias Gudmestad, Per Strass, Einar Nygård and Kenneth Johannesen for valuable input to this presentation.

**VOCABULARY**

**Baroclinic**
A baroclinic conditions is recognized by isopycnic surfaces intersects or are inclined to isobaric surfaces.

**Baroclinic current:**
Driven by vertical variations in the density of the water (isopycnics). Pressure and density levels are not coincident. The baroclinic current is varying with depth. Due to the Coriolis force the baroclinic current is forced to flow perpendicular to the pressure gradient.

**Barotropic**
A barotropic condition is recognized by isopycnic surfaces and isobaric surfaces are parallel and their slope remain constant with depth.

**Barotropic current:**
Driven by the slope of the water surface causing a horizontal pressure gradient. The barotropic currents are not varying with depth. Due to the Coriolis force the barotropic current is forced to flow perpendicular to the pressure gradient.

**Geostrophic current:**
Currents in which the horizontal pressure gradient force is balanced by the Coriolis force.

**Halocline**
The Halocline (halos = salt) is a zone of rapid salinity increase with depth. The halocline often coincides with the thermocline, and the combination produces a pronounced pycnocline.

**Isobaric surfaces:**
Surfaces of constant pressure.

**Isopycnic surfaces:**
Surfaces of constant density.

**Pendulum day**
A pendulum day is the time required for the plane of vibration of a Foucault pendulum to rotate through $2\pi$ radians. The value of one-half pendulum day is 11.97 h at the pole, 16.93 h at 45° latitude and infinity at equator.

**Pendulum hour**
$1\text{ ph} = 3590 / \sin \phi \ (\text{seconds})$
**Pycnocline**
Phycnocline (phycnos = strong) is the zone which density increases with increasing depth. This zone isolates surface water from the denser layer below (Garrison 1999).

**Sv**
1 Sv (Sverdrup) = 1 000 000 cubic meters per second

**Thermocline**
Thermocline (therm=heat, clinare=slope) is the zone between the upper layer (temperate and well-mixed surface zone) and deep layer (cold water). In this zone the temperature changes rapidly with depth (Garrison 1999).

**REFERENCES**

Apel, J. and Gonzalez, F. "Nonlinear features of internal waves off Baja California as observed from the SEASAT imaging radar", Journal of Geophysical Research, 88(C7), p 4459-4466, 1983


Bjerke, P.L. and Torsethaugen, K. "Environmental conditions on the Norwegian Continental Shelf, Barents Sea", Report no STF60 A89052, Norwegian Hydrotechnical Laboratory, SINTEF, Trondheim, 1989

Breen, Ola "Oseanografi”, Fabritius Forlagshus, 1980

Cooper, C "An overview of currents in deep water", OMAE special symposium on Offshore and Arctic Frontiers, 1986

Cummins, P. and LeBlond, P., "Analysis of internal solitary waves observed in Davis Strait", Atmos-Ocean 22, pp 173-192, 1984

Den Norske Los, Alminnelige opplysninger, Bind 1, Statens Kartverk Kystkartverket, Stavanger, 1997

Fu, L. and Holt, B., "Internal waves in the Gulf of California, observation from spaceborn radar", Journal of Geophysical Research, 89(C2), p 2053-2060, 1984

Førland, Even “Strømforhold i Nordsjøen og ellers påNorsk kontinentalsokkel”, Statoil, 1985

Gan, J. and Ingram, R., "Internal hydraulics, solitons and associated mixing in a stratified sound", Journal of Geophysical Research, 97(C6), p 9669-9688, 1992


Gjevik, Bjørn "Hva driver golfstrømmen?", Naturen, No. 4 1996

Gjevik, Bjørn "Summary and assesment of the NDP metocean project", Project report to the Norwegian Deepwater Project, June 12, 2000

Grant, C. K., Dyer, R.C., Leggett, I.M. "Development of a New Metocean Design Basis for the NW Shelf of Europe", OTC 7685, Offshore Technology Conference, 1995


Holloway, P., "Internal hydraulic jumps and solitons at a shelf break region on the Australian North West Shelf", Journal of Geophysical Research, 92(C5), pp 5405-5416, 1987


NORSOK N-003 “Action and action effects”, Rev. 1, NTS, February 1999

NPD “Regulations relating to loadbearing structures in the petroleum activities”, Norwegian Petroleum Directorate, Stavanger, 1999


Pingree, R and Mardell, G., "Solitary internal waves in the Celtic Sea", Progress in Oceanography 14, pp 431-441, 1985


Poulain, P. M. "Drifter Measurements in the Nordic and Barents Seas", International WOCE Newsletter, Number 21, December 1995, pp 19 - 21, WOCE International Project Office at Southampton Oceanography Centre UK, 1995

Roth, Jens-Christian, "Et studium av solitære indre bøger - Hovedoppgave i fysisk oseanografi", Geofysisk Institutt, University of Bergen, Norway, 2000


Sapia, A. and Salutsi, E., "Observation of nonlinear internal solitary wave trans at the northern and southern mouths of the Strait of Messina", Deep-sea research 34(7A), pp 1081-1092, 1987

Sulebak, Jan R. "Havløre", Alma Mater Publisher, Bergen 1991

