FALKLANDS CURRENT

See **BRAZIL AND FALKLANDS (MALVINAS) CURRENTS**

FIORD CIRCULATION

A. Stigebrandt, Göteborg University, Göteborg, Sweden

doi:10.1006/rwos.2001.0076

Introduction

Fiords are glacially carved oceanic intrusions into land. They are often deep and narrow with a sill in the mouth. Waters from neighboring seas and locally supplied freshwater fill up the fiords, often leading to strong stratification. During transport into and stay in the fiord, mixing processes modify the properties of imported water masses. From the top downward, the fiord water is appropriately partitioned into surface water, intermediary water and, beneath the sill level, basin water. Fiord circulation is forced both externally and internally. External forcing is provided by temporal variations of both sea level (e.g. tides) and density of the water column outside the fiord mouth. Internal forcing is provided by freshwater supply, winds, and tides in the fiord. The response of circulation and mixing in the different water masses to a certain forcing depends very much on the characteristics of the fiord topography. The circulation of basin water is critically dependent on diapycnal mixing. Hence we focus on fiord circulation from a hydrodynamic point of view. Major hydrodynamic processes and simple quantitative models of the main types of circulation are presented.

Basic Concepts in Fiord Descriptions

Some key elements of the hydrography and dynamics of fiords are shown in **Figure 1**. The surface water may have reduced salinity due to freshwater supply. It is kept locally well mixed by the wind, and the thickness is typically of the order of a few meters. The thickness of the intermediate layer, reaching down to the sill level, depends strongly on the sill depth. This layer may be thin and even missing in fiords with shallow sills. Surface and intermediate waters have free connection with the coastal area through the fiord's mouth. Basin water, the densest water in the fiord, is trapped behind the sill. It is vertically stratified, but the density varies less than in the layers above. A typical vertical distribution of density, $\rho(z)$, in a strongly stratified fiord is shown in **Figure 1**. The area (water) outside the fiord is here denoted coastal area (water).

In most fiords, temporal variations of the density of the coastal water are crucial for the water exchange, both above and below the sill level. Surface and intermediary waters in short fiords with relatively wide mouths, may be exchanged quickly (i.e. in days). The vertical stratification in the intermediary layer in such a fiord is usually quite similar to the stratification outside the fiord although there is some phase lag, with the coastal stratification leading before that in the fiord. Short residence time for water above sill level means that the pelagic ecology may change rapidly due to advection. Denser coastal water, occasionally appearing above sill level, intrudes into the fiord and sinks down along the seabed. It then forms a turbulent dense bottom current that entrains ambient water, decreasing the density and increasing the volume flow of the current (**Figure 1**). The intruding water replaces residing basin water. During so-called stagnation periods, when the density of the coastal water above sill level is less than the density of the basin water, a pycnocline develops at or just below sill depth in the fiord. In particular during extended stagnation periods, oxygen deficit and even anoxia may develop. The density of basin water decreases slowly due to turbulent mixing, transporting less dense water from above into this layer. Tides are usually the main energy source for deep-water turbulence in fiords. Baroclinic wave drag, acting on barotropic tidal flow across sills separating stratified basins, is the process controlling this energy transfer.

Major Hydrodynamic Processes in Fiord Circulation

Understanding fiord circulation requires knowledge of some basic physical oceanographic processes such as mixing and a variety of strait flows, regulating

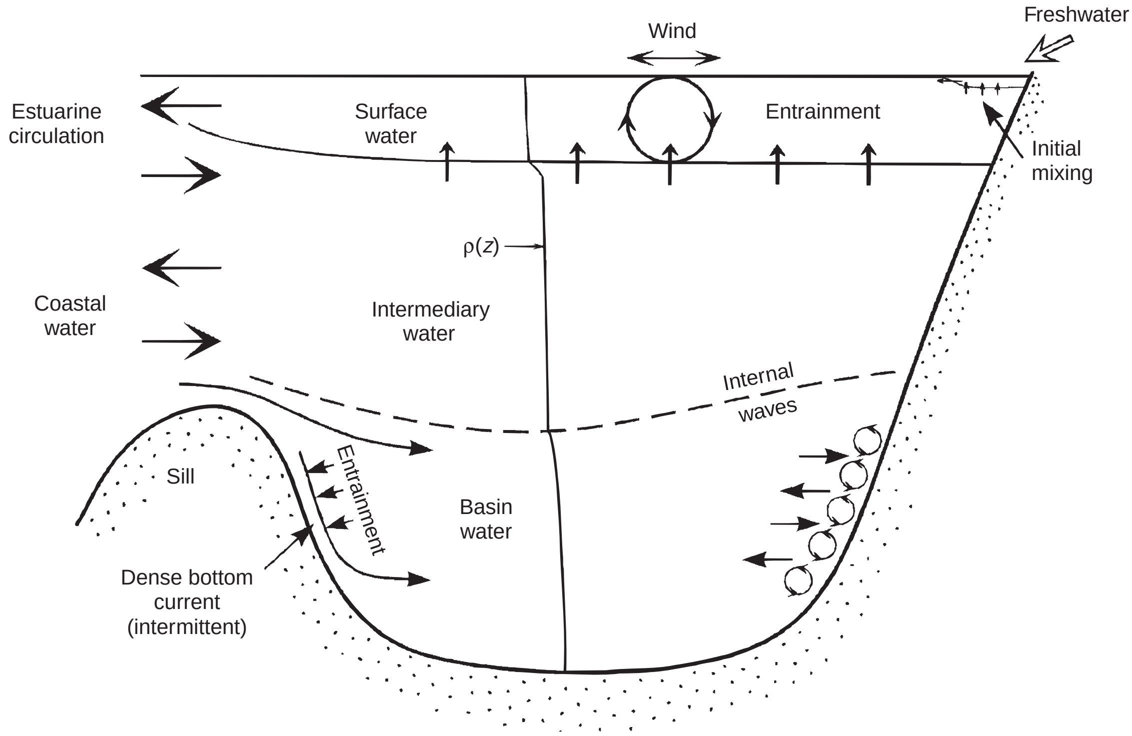

Figure 1 Basic features of fiord hydrography and circulation.

the flow through the mouth. Natural modes of oscillation (seiches) as well as sill-induced processes like baroclinic wave drag, tidal jets, and hydraulic jumps may constitute part of the fiord response to time-dependent forcing.

Flow through fiord mouths is driven mainly by barotropic and baroclinic longitudinal pressure gradients. The barotropic pressure gradient, constant from sea surface to seabed, is due to differing sea levels in the fiord and the coastal area. Baroclinic pressure gradients arise from differing vertical density distributions in the fiord and coastal area. The baroclinic pressure gradient varies with depth. Different types of flow resistance and mechanisms of hydraulic control may modify flow through a mouth. Barotropic forcing usually dominates in shallow fiord mouths, while baroclinic forcing may dominate in deeper mouths (**Figure 2**).

Three main mechanisms cause resistance to barotropic flow in straits. Firstly, friction against the seabed. Secondly, large-scale form drag, due to large-scale longitudinal variations of the vertical cross-sectional area of the strait, causing contraction followed by expansion of the flow. Thirdly, baroclinic wave drag. This is due to generation of baroclinic (internal) waves in the adjacent stratified basins.

Barotropic flow Q through a straight, rectangular, narrow, and shallow strait of width B, depth D, and length L, connecting a wide fiord and the wide coastal area, may be computed from eqn [1].

$$Q^2 = \frac{2g\Delta\eta B^2 D^2}{1 + 2C_D(L/D)} \qquad [1]$$

The sign of Q equals that of $\Delta\eta = h_o - h_i, h_o(h_i)$ is the sea level in the coastal area (fiord). This equation, contains both large-scale topographic drag and drag due to bottom friction (drag coefficient C_D).

To get first estimates of baroclinic flows and processes, it is often quite relevant to approximate a continuous stratification by a two-layer stratification, with two homogeneous layers of specified thickness and density difference $\Delta\rho$ on top of each other. The buoyancy of the less dense water is $g' = g\Delta\rho/\rho_0$, where g is the acceleration of gravity and ρ_0 is a reference density.

In stratified waters, fluctuating barotropic flows (e.g., tides) over sills are subject to baroclinic wave drag. This is important in fiords because much of the power transferred to baroclinic motions apparently ends up in deep-water turbulence. Assuming a two-layer approximation of the fiord stratification, with the pycnocline at sill depth, the barotropic to baroclinic energy transfer E_j from the jth tidal component is given by eqn [2].

$$E_j = \frac{\rho_0}{2}\omega_j^2 a_j^2 \frac{A_f^2}{A_m}\frac{H_b}{H_b + H_t} c_i \qquad [2]$$

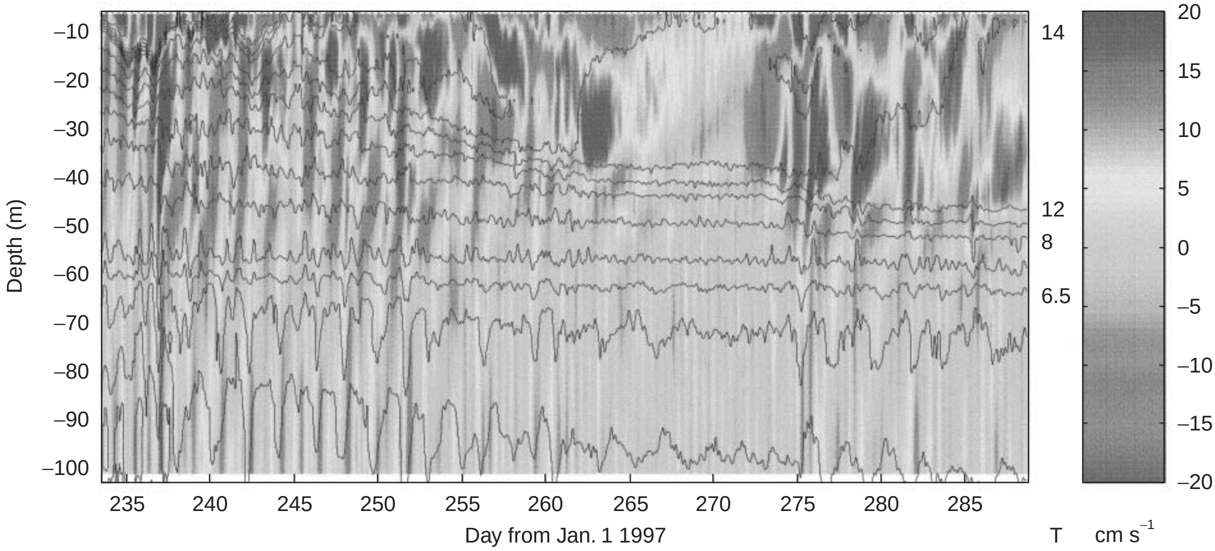

Figure 2 Time series of currents and temperature in the deepest part and about 10 km from the mouth of the Gullmar Fiord, Sweden. Currents were measured with two overlapping Acoustic Doppler Current Profilers (ADCP – RDI Instruments), one at the sea bed and one at about 40 m depth. Temperatures were measured by moored loggers. The flow is strongest above the sill level at about 40 m depth. Flow into the fiord is red and out of the fiord is blue; see the velocity scale to the right in the figure. Flow events with periods of a few days are of the intermediary type, driven by wind-driven density changes in the coastal water. Semidiurnal tides are mainly baroclinic and occur during the whole period. At times there are large temperature variations in the deepwater, below the sill level, caused by internal seiches of period 1–2 days.

Here ω_j is frequency and a_j the amplitude of the tidal component. A_f is the horizontal surface area of the fiord, A_m the vertical cross-sectional area of the mouth, $H_\mathrm{t}(H_\mathrm{b})$ sill depth (mean depth of the basin water), and $c_\mathrm{i} = [g'H_\mathrm{t}H_\mathrm{b}/(H_\mathrm{t} + H_\mathrm{b})]^{1/2}$ the speed of long internal waves in the fiord. Baroclinic wave drag occurs if the speed of the barotropic flow in the mouth is less than c_i. If the barotropic speed is higher, a tidal jet develops on the lee side of the sill together with a number of flow phenomena like internal hydraulic jumps and associated internal waves of super-tidal frequencies.

Baroclinic flows in straits may be influenced by stationary internal waves imposing a baroclinic hydraulic control. For a two-layer approximation of the stratification in the mouth the flow is hydraulically controlled when the following condition, formulated by Stommel and Farmer in 1953, is fulfilled (eqn [3]).

$$\frac{u_{1\mathrm{m}}^2}{g'H_{1\mathrm{m}}} + \frac{u_{2\mathrm{m}}^2}{g'H_{2\mathrm{m}}} = 1 \qquad [3]$$

Here $u_{1\mathrm{m}}$ ($u_{2\mathrm{m}}$) and $H_{1\mathrm{m}}$ ($H_{2\mathrm{m}}$) are speed and thickness, respectively, of the upper (lower) layer in the mouth. Equation [3] may serve as a dynamic boundary condition for fiord circulation. The model for the surface layer later in this article provides an example. It has also been applied to very large fiords like the Bothnian Bay and the Black Sea. Experiments show that superposed barotropic currents just modulate the flow. However, if the barotropic speed is greater than the speed of internal waves in the mouth, these are swept away and the baroclinic control cannot be established. In wide fiords, the rotation of the Earth may limit the width of baroclinic currents to the order of the internal Rossby radius $\sqrt{g'H_1}/f$. Here H_1 is the thickness of the upper layer in the fiord and f the Coriolis parameter. The outflow from the surface layer is then essentially geostrophically balanced and the transport is $Q_1 = g'H_1^2/2f$. This expression has been used as boundary condition for the outflow of surface water from, e.g., Kattegat and the Arctic Ocean through Fram Strait.

Diapycnal (vertical) mixing processes may modify the water masses in fiords. In the surface layer, the wind creates turbulence that homogenizes the surface layer vertically and entrains sea water from below. The rate of entrainment may be described by a vertical velocity, w_e, defined by eqn [4].

$$w_\mathrm{e} = \frac{m_0 u_*^3}{g'H_1} \qquad [4]$$

Here u_* is the friction velocity in the surface layer, linearly related to the wind speed, and m_0 (≈ 0.8) is essentially an efficiency factor, well known from seasonal pycnocline models. H_1 is the thickness of the surface layer in the fiord, and g' the buoyancy of surface water relative to the underlying water. Due to their sporadic and ephemeral character, there are no direct observations of dense bottom currents in fiords. However, from both observations and modeling of dense bottom currents in the Baltic it appears that entrainment velocity may be described by eqn [4]. For this application u_* is proportional to the current speed, H_1 is the current thickness, and g' the buoyancy of ambient water relative to the dense bottom current.

Observational evidence strongly supports the idea that diapycnal mixing in the basin water of most fiords is driven by tidal energy, released by baroclinic wave drag at sills. The details of the energy cascade, from baroclinic wave drag to small-scale turbulence, are still not properly understood. **Figure 1** leaves the impression that energy transfer to small-scale turbulence and diapycnal mixing takes place in the inner reaches of a fiord. Here internal waves, for example generated by baroclinic wave drag at the sill, are supposed to break against sloping bottoms. The temporal and spatial distributions of turbulent mixing in basin waters of fiords have not yet been mapped. However, a tracer experiment in the Oslo Fiord suggests that mixing essentially occurs along the rim of the basin, possibly close to the sill where the main part of the barotropic to baroclinic energy transfer takes place.

In a column of the basin water the mean rate of work against the buoyancy forces given by eqn [5].

$$W = W_0 + \frac{Rf\sum_{j=1}^{n} E_j}{A_t} \qquad [5]$$

Here W_0 is the nontidal energy supply, n the number of tidal components, and E_j may be obtained from eqn [2]. A_t is the horizontal surface area of the fiord at sill level and Rf the flux Richardson number, the efficiency of turbulence with respect to diapycnal mixing. Estimates from numerous fiords show that Rf equals about 0.06. Experimental evidence shows that in fiords with a tidal jet at the mouth, most of the released energy dissipates above sill level and only a small fraction contributes to mixing in the deep water.

Simple Quantitative Models of Fiord Circulation

The Surface Layer

The upper layers in fiords may be exchanged due to so-called estuarine circulation, caused by the combination of freshwater supply and vertical mixing. This is essentially a baroclinic circulation, driven by density differences between the upper layers in the fiord and coastal area, respectively. However, if the sill is very shallow, the water exchange tends to be performed by barotropic flow with alternating direction, forced by the fluctuating sea level outside the fiord.

The following equations describe the steady-state volume and salt conservation of the surface layer.

$$Q_1 = Q_2 + Q_f \qquad [6]$$

$$Q_1 S_1 = Q_2 S_2 \qquad [7]$$

Here Q_f is the freshwater supply, Q_1 (Q_2) the outflow (inflow) of surface (sea) water, and S_1 (S_2) the salinity of the surface water (sea water). Eqns [6] and [7] give eqn [8].

$$Q_1 = Q_f \frac{S_2 - S_1}{S_2} \qquad [8]$$

This equation has been used throughout the past century for diagnostic estimates of the magnitude of estuarine circulation from measurements of S_1, S_2 and Q_f.

Density varies with both salinity and temperature. In brackish waters, density (ρ) variations are often dominated by salinity (S) variations. For simplified analytical models one may take advantage of this and use the equation of state for brackish water (eqn [9]).

$$\rho = \rho_f(1 + \beta S) \qquad [9]$$

Here ρ_f is the density of fresh water and the so-called salt contraction coefficient β equals $0.0008\,S^{-1}$. The density difference $\Delta\rho$ between two homogeneous layers, with salinity difference ΔS, then equals $\rho_f\beta\Delta S$, and the buoyancy $g' = g\Delta\rho/\rho$ equals $g\beta\Delta S$. A continuous stratification in a salt-stratified system may be replaced by a dynamically equivalent two-layer stratification. This requires that the two-layer and observed stratification (i) contain the same amount of fresh water, and (ii) have the same potential energy.

Stationary Estuarine Circulation in Fiords with Deep Sills

To investigate how salinity S_1 and thickness H_1 of the surface layer depend on wind and freshwater supply, the following simple model may be illustrative. It is assumed that the fiord mouth is deep and even narrow compared with the fiord. Then the so-called compensation current into the fiord is deep and slow compared with the current of outflowing surface water. The hydraulic control condition in the mouth, eqn [3], is then simplified to $u_{1m}^2 = g'H_{1m}$. The thickness of the surface layer H_1 in the fiord is related to that in the mouth by $H_1 = \varphi H_{1m}$. Entrainment of sea water of salinity S_2 into the surface layer is described by eqn [4]. Under these assumptions, expressions for the thickness H_1 and salinity S_1 of the surface layer in the fiord may be derived (eqns [10] and [11]).

$$H_1 = \frac{G}{2Q_f g'} + \varphi\left(\frac{Q_f^2}{g'B_m^2}\right)^{1/3} \qquad [10]$$

$$S_1 = \frac{S_2 G}{G + 2\varphi\left[Q_f^5\left(\frac{g'}{B_m}\right)^2\right]^{1/3}} \qquad [11]$$

Here B_m is the width of the control section in the mouth, $G = CWs^3A_f$, $C = 2.5 \times 10^{-9}$ is an empirical constant containing, among others, the drag coefficient for air flow over the sea surface, Ws the wind speed and $g' = g\beta S_2$. Theoretically, the value of φ is expected to be in the range 1.5–1.7. Observations in fiords give φ-values in the range 1.5–2.5.

The left term in the expression for H_1 is the so-called Monin–Obukhov length, known from the theory of geophysical turbulent boundary layers with vertical buoyancy fluxes, and the right term is the freshwater thickness H_{1f}, hydraulically controlled by the mouth. The salinity of the surface layer S_1 increases with increasing wind speed and decreasing freshwater supply. For a given freshwater supply, strong winds may apparently multiply the outflow as compared with Q_f, cf. eqn [8].

The freshwater volume in the fiord is $V_f = H_{1f}A_f$. The residence time of fresh water in the fiord, $\tau_f = V_f/Q_f$, is given by eqn [12].

$$\tau_f = \varphi A_f\left(\frac{1}{g'B_m^2}\right)^{1/3} Q_f^{-1/3} \qquad [12]$$

The residence time thus decreases with the freshwater supply. It should be noted that H_{1f}, τ_f, and V_f are independent of the rate of wind mixing.

Water Exchange through Very Shallow and Narrow Mouths

In very shallow and narrow fiord mouths, barotropic flow usually dominates. The instantaneous flow which can be estimated using eqn [1], is typically unidirectional and the direction depends on the sign of $\Delta\eta$, the sea level difference across the mouth. If the mouth is extremely shallow and narrow, tides and other sea-level fluctuations in the fiord will have smaller amplitude than in a coastal area due to the choking effect of the mouth. A number of choked fiords and other semi-enclosed water bodies have been described in the literature, e.g. Framvaren and Nordaasvannet (Norway), Sechelt Inlet (Canada), the Baltic Sea, the Black Sea and tropical lagoons.

The Intermediary Layer

Density variations in the coastal water above sill level give rise to water exchange across the mouth, termed intermediary water exchange. The stratification in the fiord strives toward that in the coastal water. Intermediary circulation increases in importance with increasing sill depth. It has been found that baroclinic intermediary circulation is the dominating circulation component in a majority of Scandinavian and Baltic fiords and bays. This is probably also true in other regions, although there have been few investigations quantifying intermediary circulation.

Despite its often-dominating contribution to water exchange in fiords, the intermediary circulation has remained astonishingly anonymous and in many studies of inshore waters even completely overlooked. One obvious reason for this is that until recently a simple formula has not been available to quantify intermediary circulation. However, a rough estimate of the mean rate of intermediary water exchange, Q_i, may be estimated using a recently derived approximate expression (eqn [13]).

$$Q_i = \gamma\sqrt{B_m H_t A_f \frac{g\Delta M}{\rho}} \qquad [13]$$

Here the dimensionless empirical constant γ equals 17×10^{-4}, as estimated for Scandinavian conditions, and ΔM the standard deviation of the weight of the water column down to sill level ($\mathrm{kg\,m^{-2}}$) in the coastal water. The latter should be a hydrodynamically reasonable measure of the mean strength of the baroclinic forcing. Statistics of scattered historic hydrographic measurements may be used to compute ΔM. Equation [13] should be regarded as a precursor to a formula, yet to be developed, accommodating for the frequency dependence of ΔM.

A conservative estimate of the mean residence time for water above the sill is $\tau_i = V_i/Q_i$, where V_i is the volume of the fiord above sill level. The residence time may be shorter if other types of circulation contribute to the water exchange.

The Basin Water

In stagnation periods, the properties of the basin water change only slowly due to vertical diffusion. A rough estimate of the mean rate of diapycnal mixing in the basin water may be obtained from eqn [14].

$$\frac{d\rho}{dt} = -\frac{CW}{gH_b^2} \qquad [14]$$

Here the empirical constant C equals 2.0 and W may be obtained from eqn [5]. The vertical diffusivity κ at the level z in the basin water may be computed from the empirical expression in eqn [15].

$$\kappa(z) = \frac{W/H_b}{\rho\bar{N}^2} c_\kappa \left(\frac{N(z)}{\bar{N}}\right)^{-1.5} \qquad [15]$$

Here $\bar{N}$ is the volume-weighted vertical average of the buoyancy frequency $N(z)$ and c_κ (≈ 1) is an empirical constant.

The basin water may be expected to be completely exchanged during the period T_e defined as in eqn [16].

$$T_e = \frac{R_e}{d\rho/dt} \qquad [16]$$

Here R_e is the mean density reduction in the basin water needed to obtain a complete exchange. T_e is the residence time for basin water. Along the Norwegian West Coast, the value of R_e is about $4/3\,\mathrm{kg\,m^{-3}}$. The value of R_e should depend on the characteristics of density fluctuations in the coastal area, and may thus possibly attain different values in other regions. Combining eqns [14] and [16] shows that the residence time decreases as W increases.

Some fiords have very narrow mouths that will hamper exchange of new deep water. A measure of this is the filling time, which is the volume of the basin water divided by the flow rate of new basin water. If the filling time for a fiord is very long, the basin will be filled not only with the densest but also with less dense coastal water. The basin water in such a fiord will thus have lower density than the basin water in a neighboring fiord with similar conditions except for a much shorter filling time.

Conclusions

This essay on fiord circulation demonstrates that several oceanographic processes of general occurrence are involved in fiord circulation. Being sheltered from winds and waves, fiords are excellent large-scale laboratories for studying these processes. The mechanics of water exchange of surface and intermediary layers, as described here, should apply equally well to narrow bays lacking sills.

See also

Dispersion and Diffusion in the Deep Ocean. Estuarine Circulation. Flows in Straits and Channels. Internal Tidal Mixing. Internal Tides.

Further Reading

Aure J, Molvær J and Stigebrandt A (1997) Observations of inshore water exchange forced by fluctuating offshore density field. *Marine Pollution Bulletin* 33: 112–119.

Freeland HJ, Farmer DM and Levings CD (eds) (1980) *Fjord Oceanography*. New York: Plenum.

Farmer DM and Freeland HJ (1983) The physical oceanography of fiords. *Progress in Oceanography* 12: 147–220.

Stigebrandt A (1999) Resistance to barotropic tidal flow by baroclinic wave drag. *Journal of Physical Oceanography* 29: 191–197.

Stigebrandt A and Aure J (1989) Vertical mixing in the basin waters of fiords. *Journal of Physical Oceanography* 19: 917–926.

FIORDIC ECOSYSTEMS

K. S. Tande, Norwegian College of Fishery Science, Tromsø, Norway

doi:10.1006/rwos.2001.0092

Introduction

Fiords and semienclosed marine systems are characterized by distinct vertical gradients in environmental factors such as salinity, temperature, nutrients, and oxygen; sampling of biological variables is considerably less adversely influenced by currents and weather conditions than in offshore systems. Therefore, the fiords are particularly well-suited for detailed studies of the pelagic habitat and have traditionally attracted marine scientists, mainly as a site for curiosity-driven research. They are easily accessible and have therefore been used as experimental laboratories for testing new