DISPERSION IN SHALLOW SEAS

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Introduction

The study of marine dispersion is particularly, but not exclusively, concerned with understanding and predicting the fate and impact of pollutants, both from acute spills from shipping and coastal facilities and from longer-term (chronic) discharges. Marine pollutants are primarily of anthropogenic origin and so most significantly affect the regions of the marine environment closest to centres of human activity, namely, estuaries, bays, and the shallow seas of the continental shelves. Some pollutants may be regarded as natural, such as toxic algal blooms; these also primarily impact human activity, such as recreation and fisheries, in shallow seas.

The environmental impact of pollutants depends on extremely complicated biogeochemistry and ecotoxicology; however, in almost every instance the effect depends in some way on the amount of contaminant present. Hence an appreciation of how pollutants are transported, dispersed, and diluted in shallow seas is crucial to our understanding of their impact. This draws on all aspects of shallow sea physical oceanography and our aim here is to assess the processes that determine the horizontal dispersion of a patch of contaminant. As with many problems in oceanography, it is difficult to define a completely general set of significant processes, as they will always depend on the particular situation at hand, so we will illustrate the general principles with a number of applications to specific cases.

Fundamentals – The Fluid Mechanics of Dispersion

For most practical purposes in shallow sea studies the evolution of the ensemble mean concentration, $\bar{C}(x, y, z, t)$, of some constituent of the water in the velocity field, $\mathbf{u} = (u, v, w)$, can be represented by the advection-diffusion equation [1],

$$\frac{\partial \bar{C}}{\partial t} + \bar{\mathbf{u}} \cdot \nabla \bar{C} = \nabla \cdot (\mathbf{K} \nabla \bar{C}) + S \qquad [1]$$

where S represents any sources/sinks present.

Instantaneous values have been divided into mean and fluctuating components as in eqn [2].

$$\mathbf{u} = \bar{\mathbf{u}} + \mathbf{u}', \qquad C = \bar{C} + C' \qquad [2]$$

Turbulent fluxes are taken to be proportional to spatial gradients of mean quantities, for example, as in eqn [3].

$$\overline{u'C'} = -K_x \frac{\partial \bar{C}}{\partial x} \qquad [3]$$

$\mathbf{K} = (K_x, K_y, K_z)$ is the turbulent diffusivity; to appreciate the significance of this quantity it is necessary to consider the processes it represents in some detail.

The only way to change the constituent properties of an infinitesimal water parcel is through molecular diffusion. The typical scale of this process is $\nu \sim 10^{-6}\,\mathrm{m^2\,s^{-1}}$, so diffusion across a 100 km wide shelf sea would take $O(10^8)$ years if it were the only active process. However, both turbulent and mean flows can reduce this timescale to the extent that they are the only significant processes. Currents will transport water masses and their constituents around shelf seas; however, it is through the process of straining (stretching) the surfaces that separate bodies of water with different constituents that dispersion in shallow seas occurs; an example of such a surface might be the interface between water containing a particular pollutant and water in which that pollutant is absent.

Horizontal variations in velocity (shears) can be expressed as the sum of a solid body rotation and a pure strain. As long as the shear persists, this strain field will tend to lengthen material contours, for example at the interfaces described above. This increases the surface area, and reduces the interface width, with a consequent increase in the diffusive flux between the water masses. This process is particularly evident in turbulent flows, where the randomly oriented strain field leads to a continual stretching and thinning of the surfaces between water masses; these quickly become very sinuous, folded, and complicated, and the diffusivity effective on a macro scale, **K**, is increased by many orders of magnitude above the molecular value, ν. There are a number of processes in the shelf sea environment that act to generate these shear and strain fields, some of which are now described.

Turbulence and Eddies

The dominant process that generates the strain fields described above is, directly or indirectly, turbulence. This can be characterized by a length scale, or wavenumber k, according to the theories of Kolmogorov. The statistical characteristics of the smallest scales of motion (the viscous range) are determined by the rate of dissipation of kinetic energy (ε) and the kinematic viscosity (v), which define the typical dimensions (L_v), life times (T_v) and velocities (U_v) associated with the smallest turbulent vortices:

$$L_v = \left(\frac{v^3}{\varepsilon}\right)^{1/4} \qquad T_v = \left(\frac{v}{\varepsilon}\right)^{1/2} \qquad U_v = (\varepsilon v)^{1/4} \quad [4]$$

Similarity theory requires the wavenumber spectrum of turbulent kinetic energy to have the form

$$E(k) = (\varepsilon v)^{1/4} F(L_v k) \quad [5]$$

where $F(L_v k)$ is a universal function. In the range of scales where viscosity is negligible but $1/k$ is still much smaller than the bulk flow macro scale, L (i.e., in the inertial subrange: $L_v \ll 1/k \ll L$), the energy spectrum must be independent of viscosity and depend only on the transfer of energy from larger to smaller scales. This leads to the $-5/3$ power law of Kolmogorov and Obukhov (eqn [6]).

$$E(K) \sim \varepsilon^{2/3} k^{-5/3} \quad [6]$$

If it is assumed that dispersion of a patch (of, say, a contaminant) of size l is mainly due to eddies also of size l, then the eddy diffusivity for this patch must be as given by eqn [7] as long as l is within the inertial range.

$$k \sim \varepsilon^{1/3} l^{4/3} \quad [7]$$

This similarity argument is only valid for fully developed three-dimensional turbulence, so L is limited either by the water depth or by a stratification scale, since eddies larger than this will inevitably be quasi-two-dimensional.

Two-dimensional flows have the property, arising from the conservation of vorticity, that the energy cascade is reversed and small scales cascade to large (again obeying a $k^{-5/3}$ law). This results in the comparatively stable and long-lived meso-scale eddies prevalent in the open ocean (the oceanic analogue of atmospheric weather systems). The frictional nature of shelf seas, however, generally limits the occurrence of these large-scale two-dimensional eddies to deeper or stably stratified regions (see later). The dispersion of tracer patches in the North Sea, for example, does not reflect the behavior expected of a field of horizontal eddies alone, but rather a complex interaction between vertical and horizontal shear and mixing processes as described next.

Tidal Currents and Shear Dispersion

At first sight the oscillatory nature of tidal currents makes them an unlikely candidate for a dispersive process. However, a conundrum found in tidal waters is the disparity between estimates of horizontal diffusion coefficients based on tracer releases ($K_x \sim 100$–$1000\,\mathrm{m^2\,s^{-1}}$) and those calculated from the typical horizontal eddy velocities and length scales in these waters ($K_x \sim 0.1$–$1\mathrm{m^2\,s^{-1}}$): tidal waters can be strongly dispersive but eddies alone are insufficient to explain this. While the answer to this discrepancy depends on the particular flow field and geometry in question, there are a number of processes that contribute to the enhanced lateral dispersion in tidal waters. One is an interaction between horizontal shear and the horizontal dispersion generated by vertical shear. This is essentially the shear-straining mechanism for enhancing diffusivities described above. The tidal flow over the seabed generates a vertical shear, which in turn generates turbulence, the diffusion coefficient of which may be written as $K_z = C_D U H$ (where $C_D \sim 0.005$ is a drag co-efficient; U is a tidal velocity scale; H is a vertical length scale, say, the water depth), although there are many other forms for this. This turbulence is strained by the vertical shear to give an enhanced horizontal diffusion; the upper limit of this (for the case of a mixing timescale much smaller than the tidal period) is given by eqn [8].

$$K_x \approx \frac{U^2 H^2}{240 K_z} \sim 170 K_z \quad [8]$$

This in itself is insufficient to account for the values of K_x quoted above; however, if this dispersion is in turn strained by a lateral shear, values of the required magnitude can be reached. Of course the nature of the horizontal shear depends on the particular situation, but as an example a unidirectional residual current, U_0, with a sinusoidal across-stream profile (length scale D) gives an enhanced horizontal diffusivity as in eqn [9].

$$K_{xx} = \frac{U_0^2 D^2}{16\pi^2 K_x} \quad [9]$$

For values $U_0 = 0.1\,\mathrm{m\,s^{-1}}$, $D = 5\,\mathrm{km}$, and $K_x = 10\,\mathrm{m^2\,s^{-1}}$ this gives $K_{xx} = 160\,\mathrm{m^2\,s^{-1}}$. Two important points to note here are, first, that a steady horizontal shear is much more efficient at dispersion than an oscillatory one; and, second, that when a uniform oscillatory current and a sheared residual current are combined, underlying turbulence is not necessarily required for dispersion (as it is in the above argument). Scales of flow can be chosen that lead to two particles that are initially close together in position to diverge rapidly even in the absence of turbulence – an example of deterministic Lagrangian chaotic behavior.

Wind-driven Currents

Away from regions of strong density variation, wind stress drives the dominant residual currents in shelf seas, and since these have large horizontal scales and can be considered quasi-stationary as far as the above analysis is concerned, their horizontal variations contribute strongly to shear-dispersion. Persistent basin-wide jetlike circulations such as the Dooley current from the northern North Sea into the Skagerrak have widths of $O(100\,\mathrm{km})$, and hence have very high effective diffusivities at their periphery and disperse material over large distances.

Wind stress also contributes to vertical shears and mixing. In combination with the Coriolis force it leads to currents that decrease from the surface value, oriented at 45° right (left) of the wind direction, and turn to the right (left) in the Northern (Southern) Hemispheres: the Ekman spiral. The turbulence resulting from this shear can create a mixed layer in thermally stratified conditions. This mixed layer is bounded at the bottom by a thermocline that isolates the surface from deeper waters and prevents vertical mixing of contaminants. The reduction in vertical mixing by stratification can enhance the transport and dispersion of material in the surface layer as this material is confined to the region where the wind-driven currents are strongest. In regions that stratify seasonally, this suggests that a given wind speed can result in greater dispersion of a surface contaminant in summer than in winter. However, this effect is likely to be more than compensated for by winds generally being stronger in winter than in summer.

While much of the focus of wind effects on dispersion is on the generation of large-scale currents, small-scale turbulence, and surface waves, the interaction of wind stress and surface waves can also result in rows of vortices of alternating sign known as Langmuir circulation. These are in the vertical plane and have axes aligned with the wind direction. These structures manifest themselves, through convergent surface currents, as rows of floating material oriented in the wind direction, called windrows. While it is known that they are associated with enhanced currents in the wind direction and very strong downward currents underneath the wind rows, their role in surface mixing and horizontal transport is still something of a mystery.

Baroclinic Processes

Density-driven currents are common in shelf seas; they arise as a result of horizontal variations in pressure due to variations in the density field. Examples include jetlike features at tidal mixing fronts where well-mixed water meets thermally stratified water, and coastal currents due to strong salinity variations such as in river plumes. The characteristic horizontal scale of these features is the Rossby radius of deformation, $R = Nh/f$ (where f is the Coriolis parameter, N the buoyancy frequency and h a vertical length scale). These currents can be unstable and lead to the formation of (quasi) two-dimensional meso-scale eddies (for example, in the Norwegian Coastal Current); however, the role of these eddies in dispersion across shelf seas is not well established. Since material circulating in closed orbits does not disperse, a persistent gyre circulation at a fixed location (such as seen around the bottom cold water dome in the western Irish Sea) will tend to restrict the dispersion of material. In contrast a meso-scale eddy detached from a coastal current will transport material large distances and disperse it as the eddy dissipates. The relation between the motion around closed orbits (u_0) and the internal diffusion velocity scale (K_x/L) within an eddy is the Peclet number: $Pe = u_0 L/K_x$. With the typical values: $u_0 = 0.5\,\mathrm{m\,s^{-1}}$, $L \sim R \sim 5\,\mathrm{km}$ and $K_x = 10\,\mathrm{m^2\,s^{-1}}$, we find $Pe \sim 250$, which means that material makes many closed orbits before diffusing. Because of their longevity and their ability to transport material as a coherent structure, two-dimensional meso-scale eddies are not well represented by diffusion coefficients, but to give an example of the magnitude of their effect, we can estimate $K_{xx} \sim u_0 L\, Pe^{-1/2}$ with the values given above, this gives $K_{xx} \sim 160\,\mathrm{m^2\,s^{-1}}$ (cf. the horizontal shear-dispersion value calculated above). As well as generating meso-scale eddies, baroclinic currents, such as those seen around cold domes, will contribute to the horizontal shear-dispersion described in the last section, particularly because they are often very localized, for example, to a narrow frontal region, and as such are associated with large horizontal gradients (they are often referred to as frontal jet) and so can result in small regions of very high horizontal diffusivity.

Surface Waves

Away from the near shore zone, wind-generated surface waves are generally non-dispersive. Their only mechanism for net transport is by Stokes drift. This results from the orbital motion of particles in the wave's velocity field being helical (rather than elliptical) because the velocity decreases with depth and the forward velocity of a particle near the surface is greater than its reverse velocity at the bottom of the orbit. This leads to a net flow U_S in the direction of wave propagation given by eqn [10] for deep water waves of amplitude A.

$$U_S = A^2 g^{1/2} k^{3/2} \exp(2kz) \qquad [10]$$

This can be significant to the transport and dispersion of small-scale patches such as oil slicks (see below).

Dispersion Phenomena

There have been numerous tracer, dye release, and drifter experiments, along with theoretical analyses and numerical experiments to examine the many phenomena in which these dispersion processes play a significant role.

Passive Dispersion

Passive tracers provide the ideal tool for studying dispersion processes. Dyes and radioactive tracers have proved most useful in marine studies to map advection and diffusion in shelf seas over timescales ranging from minutes to many years. Most notable has been the release of ^{137}Cs (with its half-life of 30.1 years) from the Sellafield (Windscale) nuclear reprocessing facility on the west coast of England during the 1970s and 1980s. This tracer, well mixed in the water column, has been observed to travel around northern Scotland, taking 2 years to reach into the North Sea. This has enabled marine scientists to estimate the contributions of advection and horizontal diffusion to the changing spatial distribution. Model studies that take into account advective and dispersive effects have produced quantitatively good agreement with the observed distributions over 15-year simulations with horizontal dispersion coefficients proportional to αR^2, where α is a constant value and R is the local tidal current amplitude.

Oil Slicks

When crude oil or petroleum products are released into the sea they are immediately subjected to a variety of degradation and dispersion processes, including advection, turbulent diffusion, spreading, evaporation, emulsification, dissolution, photochemical oxidation, aerosol formation, sedimentation, and biodegradation. The composition and position of release of the oil greatly influence the relative importance of these processes. Certain refined products like petroleum and light kerosenes may be subject to almost complete evaporation; heavier oils do not undergo significant weathering and are more likely remain in the water and therefore be subject to the advection–diffusion processes arising from tides, wind, and wave action.

If a slick is introduced into a uniform current then as it diffuses and increases in size it will be advected by the current without any distortion of the patch slick shape. However, if the current is spatially nonuniform (horizontally and vertically) then the slick will be distorted and elongated and display apparent enhanced diffusion in the direction of the dominant current (**Figure 1**). The elongation of the oil slick in the direction of the wind and waves suggests the presence of a shear-diffusion process. Vertical shears in the surface water are produced by the action of wind and waves and these processes can also mix oil droplets down into the water column, where they are diffused by the shear. The interaction between the turbulent diffusion of oil droplets and the vertical shear beneath the slick determines the growth dimensions of the slick, with shear diffusion elongating the slick in the direction of the wind/waves and Fickian diffusion affecting the slick width. Tides will act to move the slick

Figure 1 Dye diffusion experiment, showing horizontal elongation of dye patch.

backward and forward but will contribute little (except in very shallow water with strong tidal currents) to the diffusion process. Typically it is considered that the wind-driven surface current moves at about 2–3.5% of the wind speed and the surface wave drift (Stokes drift) has a velocity of about 1–2% of the wind speed. However, for typical wave conditions the component of the shear diffusion will only be about 10% of that due to the wind-induced flow.

Plankton Patchiness

The vast majority of marine organisms are planktonic and are thus largely at the mercy of the motions in the sea. It has been known for many years that plankton are neither randomly nor uniformly distributed in the sea; rather they exist in patches, or exhibit patchiness. It is still not clear in which situations the spatial distribution of plankton is controlled by physical processes (dispersion/concentration) and in which it is controlled by biological ones (growth/reproduction/behavior).

The spectra of chlorophyll patches (of size L) in the sea have been shown to follow the $k^{-5/3}$ shape described above (here $k \sim 1/L$). While spectra for higher trophic levels (e.g., zooplankton) seem to be less steep than k^{-2} at low wavenumbers, in general many biological quantities have spatial distributions that appear to be determined solely by the physics.

It has, however, been shown that biological interaction can modify the spectral distribution shape. In particular, spectral shape is modified by interaction between species. If there is no interaction in the population then, at length scales smaller than the 'kiss length' (the minimum size patch that can maintain itself in the presence of diffusion), the distribution follows the $k^{-5/3}$ distribution (i.e., solely controlled by the physical turbulence), whereas at larger scales the spectrum is proportional to k^{-1} (i.e., flatter ('whiter') than the physical turbulence). Thus, small-scale plankton structures appear totally controlled by turbulence, whereas large-scale structures show less variance than the underlying physical structures. When species interact (e.g., predator–prey interaction) in the inertial sub-range the interaction produces a k^{-3} spectral shape; that is, there is increased variance at the larger length scales. However, if the underlying turbulence is fully two-dimensional (enstrophy-conserving) then the converse is true: the population interaction leads to less (k^{-1}) intense patchiness at larger scales (**Figure 2**). Thus biological interaction can either redden (increase) or whiten (decrease) the spectral variability of patchiness due to physical diffusion alone,

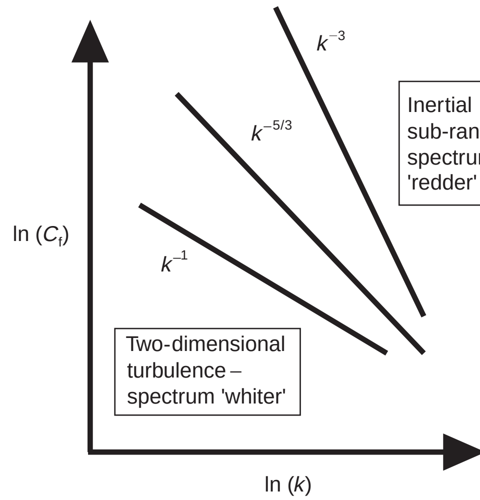

Figure 2 Schematic of power spectrum of concentration fluctuation (C_f) against wavenumber (k) for two interacting species in different turbulence regimes. The line $k^{-5/3}$ represents the spectrum of physical turbulence.

depending on the turbulent length scale. A similar result is seen in terrestrial environments. In simple terms, at low turbulence levels (too low to disperse patches) predator–prey encounter is increased; at intermediate levels patches are dispersed, making predator–prey encounter difficult; at high levels the prey distribution is homogenized, resulting in increased predator–prey encounter rates.

It has, however, been suggested that the diffusion approach to plankton patchiness is too readily applied and that it neglects the fact that true diffusion within the surface layers of the sea is insignificant at larger than centimeter scales. It has been shown that stirring by a turbulent flow causes variability to be transferred from large to small scales and that under the influence of turbulent advection a patch of tracer (e.g., phytoplankton) develops fine tendrils and filaments (as described above). In simple models of coupled phytoplankton and zooplankton, fine structure in the zooplankton distributions can be generated by the transfer of variability from larger scales by stirring over the lifetime of zooplankton. The timescale at mid-latitudes for the transfer of variance from the large (100 km) scale motions to the small (1 km) scale is typically 10 days, which is less than the lifetime of larger zooplankton such as copepods (typically 25 days to reach maturity). Thus any large-scale variation in juvenile copepod distribution will be stirred down to kilometer length scales. Spectral shapes resulting from these studies show the zooplankton to have a flatter spectra than the phytoplankton with exponents representative of fully two-dimensional turbulent flow and the experiments indicate that zooplankton lifetime is an important determinant in their spatial pattern.

Suspended Particulate Matter Dispersion

The transport of the organic and inorganic suspended material typically found in sea water represents a somewhat different class of dispersion phenomena important for a number of reasons: its role in biogeochemical cycling and the optical properties of sea water important for biological production; the transport of pollutants that preferentially adhere to particles (such as many heavy metals); and in the long term because of its implications to bed forms and coastal morphology. This material, collectively referred to as suspended particulate matter (SPM), will tend to settle out of the water column and deposit on the seabed. Its horizontal transport is crucially dependent on the rate of settling (a factor of the particle size), and the ability of currents (due to wind, waves and tides) to scour the seabed and erode/resuspend the material back into the water column.

Coarse material (grain sizes larger than $\sim 0.1\,\mathrm{mm}$) generally only moves as bed load or during strong storm events and, because of its short residence time in the water column, is primarily involved only in benthic processes, bedforms, and coastal morphology, rather than pollutant transport, or optical properties. In contrast fine material, with settling velocities up to about $10\,\mathrm{mm\,s^{-1}}$ (grain size of $\sim 0.1\,\mathrm{mm}$) can be treated in a similar fashion to dissolved tracers with the addition of settling, resuspension, and deposition terms to the advection–diffusion equation. The former is simply vertical advection at the settling velocity, while empirical forms for the other terms must be found. These often relate the rates of erosion and deposition to critical bed stresses $(\tau_{\mathrm{ero}}, \tau_{\mathrm{dep}})$, for example, as in eqns [11], where B is the amount of deposited material (m^{-2}), w is the settling velocity, and M is an empirical erosion rate.

$$\frac{\partial B}{\partial t} = -M(\tau/\tau_{\mathrm{ero}} - 1), \quad \tau > \tau_{\mathrm{ero}}$$

$$\frac{\partial B}{\partial t} = -wC|_{z=-H}(\tau/\tau_{\mathrm{dep}} - 1), \quad \tau < \tau_{\mathrm{dep}} \qquad [11]$$

This can have a marked effect on the dispersion and transport, since under conditions of weak bed stress (for example, neap tides) the material remains on the seabed and is not transported (unlike dissolved material in the water column). This can lead to the transport being dependent on correlations between the tidal cycles, and variations in the wind and wave climates. There is a tendency for material to settle out in the summer month and be resuspended during winter storms. This seasonality can have a marked effect on the transport (compared with dissolved material) if winter and summer currents differ significantly, for example, owing to density driven currents and variation in the wind forcing.

See also

Ekman Transport and Pumping. Langmuir Circulation and Instability. Meddies and Sub-surface Eddies. Mesoscale Eddies. Oil Pollution. Patch Dynamics. Surface, Gravity and Capillary Waves. Wind Driven Circulation.

Further Reading

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