Return to previous file: MCU6a

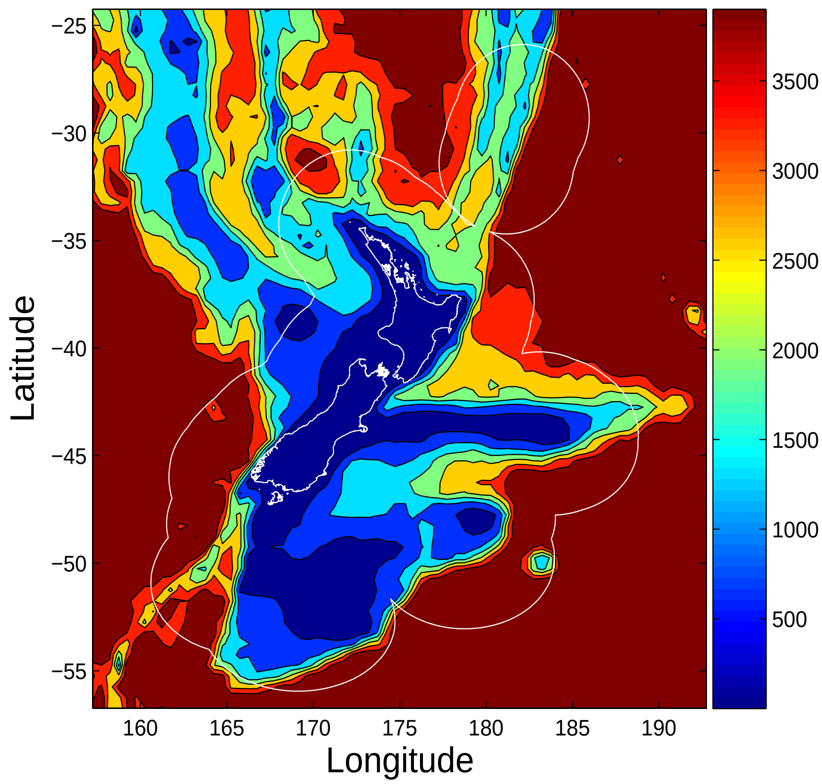

Figure 3.9: Contour plot of either the estimated lysocline depth ELD or the bottom depth, whichever is shallower, for the NZ oceanic region. The NZ coastline and EEZ boundaries are shown in white.

3.14 Interpolation to Output Grid

Data at ½° resolution were interpolated to the MEC Mercator output grid at NIWA by M. Pinkerton (NIWA Wellington), and as ESRI grid files by Ude Shankar (NIWA, Christchurch).

4. Photosynthetically Active Radiation (PAR) at the Sea-bed

4.1 Introduction

The intensity of broad-band visible incident irradiance at the sea-bed depends in a complex way on the spatial and temporal distribution of light at the sea-surface, and on the optical properties of the water column. In waters with a vertically homogenous distribution of coloured material, downwelling irradiance decreases approximately exponentially with depth. The diffuse attenuation coefficient for natural oceanic waters is likely to be greater than 0.05 m^{-1}. We can hence estimate that for waters deeper than 200 m, the light at the sea-bed will be less than 0.00005 of the surface value. Most of the interest for the Marine Environment Classification will hence be for shallow coastal and shelf waters round New Zealand, and we pay particular attention to these regions in the work described below.

The value of the product "PAR downwelling irradiance at the sea-bed" depends on what it will be used for. This product will give a good general indication of the amount of light available for photosynthesis by benthic phyta, including macroalgae (kelp etc), sea-grass, and microphytobenthos. However, that there are other measures of underwater light intensity that may be more appropriate to particular classes of benthic phyta. For example, kelp growing over a reflective sandy-bottom may also receive light reflected upwards from the sand. Light available for microphytobenthos on a rock will depend on the inclination and orientation of the rock relative to the sun. Red algae and green algae have different action spectra (they absorb different colours of light for photosynthesis), so a broad-band average (PAR) may not be the ideal measure of light intensity for all groups. Currently, these issues can only be resolved at small scales where we have detailed information on the type of benthic phyta.

4.2 Method

At any instant of time, the downwelling irradiance at the sea-bed can be estimated by Equation [4.1]. We assume a vertically homogenous water column. This assumption is discussed further later. Reflectance of light from the sea-bed is a second-order effect, and is neglected.

$$E_{seabed}(\lambda, d, T, x, y) = E_{0-}(\lambda, d, T, x, y) \cdot \exp[-K_E(\lambda, d, x, y) \cdot Z(x, y)] \qquad [4.1]$$

E_{seabed} = downwelling irradiance at sea-bed (W m^{-2} nm^{-1})
E_{0-} = downwelling irradiance just below sea-surface (W m^{-2} nm^{-1})
λ = wavelength (nm)

d = day of year
T = time of day
x, y = location
Z = water depth (m)
K_E = diffuse downwelling irradiance attenuation coefficient (m^{-1})

Photosynthetically-available radiation (PAR) at the sea-bed is related to spectral irradiance as Equation [4.2]. We use the standard wavelength limits of 400–700 nm as our definition of PAR.

$$PAR(d,T,x,y) = \frac{1}{N_a} \cdot \frac{1}{hc} \cdot \int_{400}^{700} E(\lambda,d,T,x,y) \cdot \lambda \cdot d\lambda \qquad [4.2]$$

PAR = Photosynthetically-available radiation (Einsteins m^{-2} s^{-1})
h = Planck's constant (6.63×10^{-34} Js)
c = speed of light ($\sim 3 \times 10^{8}$ ms^{-1} in air)
N_a = Avogadro's constant (6.022×10^{23} mol^{-1})

The long-term average light at the sea-bed at a given point is given as the integral of Equation [4.1] for each time of day, and day. A complete solution would hence require spectral estimates of incident light, and diffuse attenuation for each location, date, and time-of-day. These data are generally not available. For example, properties of the water column are generally estimated only 1–3 times daily via satellite observations, and the effect of clouds and sea-surface roughness on the irradiance just below the sea-surface at high spatial and temporal resolution are not known.

Here, we made the simplifying assumption that PAR at the sea-bed is largely determined by two factors that are largely independent over time scales of ~8 days: (1) total incident PAR just below the sea-surface; (2) attenuation of light between the sea-surface and the sea-bed. By assuming that these factors are independent at such time scales, it is possible to estimate the long-term average PAR at the sea-bed. The assumption that the variation in the spectral composition of downwelling irradiance incident on the sea-surface over short time scales (minutes for clouds, hours for solar zenith angle) has a small effect on PAR at the sea-bed is reasonable. Most PAR at the sea-bed will be due to light in the wavelength range 450–550 nm, and in this range, extraterrestrial solar irradiance, and effects such as the reflection of incident light by clouds, absorption by water-vapour and ozone, and the refractive index of water (which effects penetration of light through the air-water interface) are approximately independent of wavelength. At extreme solar zenith angles (→90°) where the spectral variation in incident irradiance may be significant, total irradiance is generally low, and the contribution to the long-term average PAR at the sea-bed will be small.

We hence divided the year into 46 periods of (mostly) 8 days, representing 1–8 January, 9–16 January and so on. The 46th period is only 5 days long. For each of these periods, for each 4 km pixel in the New Zealand region, we calculated an average, daily incident PAR just below the sea-surface, and the average diffuse downwelling PAR attenuation coefficient. These were combined as Equation [4.3] to give the climatological sea-bed PAR value for each pixel, PAR_{out}(x,y).

$$PAR_{out}(x,y) = \sum_{i=0}^{45} PAR_{0-}^{i}(x,y) \cdot \exp\left[-K_E^i(x,y) \cdot Z(x,y)\right] \cdot \frac{T^i}{365} \qquad [4.3]$$

T^i = Length of interval (generally 8 d except for 46th interval)

4.3 Bathymetry

The depth at each pixel was taken from topography dataset generated at NIWA and used in the generation of Layer 1: "Temperature at the sea-bed" as explained in Section 2 of this report.

4.4 Estimation of Incident Irradiance

PAR just below the sea-surface was estimated as follows. First, we estimate mean extra-terrestrial solar irradiance, and correct for Earth-Sun distance effects through the year. This top-of-atmosphere irradiance is propagated through the atmosphere to the sea-surface accounting for cloud effects, absorption by water vapour, oxygen and ozone in the atmosphere, and for backscattering by atmospheric molecules and aerosols as Equation [4.4].

$$E_{0+} = F_0(\lambda) \cdot D(d) \cdot t(\theta_0, \lambda) \cdot C(d, x, y) \qquad [4.4]$$

E_{0+} = downwelling irradiance just above sea-surface (W m^{-2} nm^{-1})
F_0 = mean extraterrestrial solar irradiance (W m^{-2} nm^{-1})
D = Earth-Sun distance correction factor (dimensionless)
t = atmospheric diffuse transmittance (dimensionless)
C = fractional reduction in incident irradiance due to cloud (dimensionless)
θ_0 = solar zenith angle

In this study, we took mean extraterrestrial solar irradiance from Wehrli (1985). The Earth-Sun distance factor correction is taken from Spencer (1971). Declination of the Earth is calculated following Kirk (1994). Absorption in the atmosphere is principally due to atmospheric water vapour (H_2O), oxygen (O_2) and ozone (O_3). The atmospheric transmittance model used here for a cloudless atmosphere follows Bird (1984). Water vapour, ozone and oxygen absorption followed Leckner (1978).

Incident irradiance was integrated between dawn and dusk (θ_0 =0°) to give the total daily insolation. This is integrated by wavelength as Equation [4.2] to give incident PAR. As a first approximation, and discounting the effects of clouds and variations in atmospheric transmittance for the time being, daily PAR insolation is a function of date and latitude only. The effect of clouds on incident irradiance at the sea surface is considered below.

4.5 Reflectance of Water Surface

A proportion of light incident on the sea-surface is reflected, and a proportion is transmitted through the interface. The proportion of incident irradiance entering the sea follows Snell's Law and Fresnel's Equation (e.g. see Kirk 1994) and is dependent on the angular composition of incident irradiance and on the amount of wind-roughening of the surface of the sea. The angular composition of incident irradiance is affected by the proportion of skylight to total sea surface irradiance and this will vary with the amount and type of cloud cover, humidity, aerosol properties and solar elevation.

In this study, we estimate downwelling irradiance just below the sea-surface, E_{0-}, as Equation [4.5], accounting for the diffuse reflectance of light from the water surface.

$$E_{0-} = E_{0+} \cdot \left[1 - R(\theta_0)\right] \qquad [4.5]$$

R = surface reflectance factor for direct solar beam and diffuse skylight (dimensionless)

We assume that diffuse reflectance of the sea-surface (R) is spectrally neutral, and may be parameterised in terms of the solar zenith angle only. We assume an overcast sky, and a typical wind-speed of 4 ms^{-1} (~8 kts). Following published information, we assume that R varies from about 0.052 for θ_0<50° (Kirk 1994; Preisendorfer 1957; Jerlov 1976) rising to about 0.20 as $\theta_0 \rightarrow 90°$.

4.6 Effect of clouds

The effect of clouds on incident irradiance at the sea-surface is complex, depending on type, distribution, cloud height, and solar zenith angle. The effect of clouds on sea-surface irradiance around New Zealand is poorly known. In this work, we estimated the long-term effects of clouds on sea-surface light as follows. First, we assumed that the diminution of irradiance incident on the sea-surface by clouds is spectrally invariant. Second, we estimate that proportion of the time clouds are present at a given point in the New Zealand domain for the long time-series of data from the SeaWiFS ocean colour satellite sensor. This was calculated for 46 periods of 8 days. Under a thin sheet of cirrus cloud, Monteith (1973) suggests that total irradiance may be about 70% of that under a clear sky. A deep layer of stratus cloud may transmit only 10% of the solar radiation, the majority of the incident irradiance being reflected back into space from the upper surface of the cloud, and the remainder being absorbed within the cloud itself (Kirk 1994). On a day with broken cloud, the instantaneous incident irradiance may be between perhaps 20–50% of the clear-sky value. Monteith (1973) states that average daily insolation is typically 50–80% of that on cloudless days. Here, we assume that on average, the daily insolation is reduced to 65% of the clear sky value whenever a pixel is observed as being cloud covered. E_{0-} can then be converted to PAR_{0-} as Equation [4.2].

4.7 Underwater Diffuse Downwelling Attenuation Coefficient

Diffuse attenuation was estimated for each location and date using data from the SeaWiFS satellite ocean colour sensor via the proxy of near-surface chlorophyll-a concentration. Chlorophyll-a concentration was estimated for the New Zealand region using the SeaWiFS OC4v4 algorithm under SeaDAS 4.7 (O'Reilly et al. 2000; Fu et al. 1998). This has been shown to be accurate to within approximately 30% in the New Zealand region, in open ocean water (Pinkerton et al. in press). In coastal waters containing appreciable concentrations of total particulate material (>0.5 g m^{-3}) and/or significant coloured dissolved organic matter (absorption by dissolved matter at 380 nm of >0.1 m^{-1}), chlorophyll concentrations estimated by the OC4v4 algorithm will not be accurate. However, preliminary work in New Zealand waters suggests that this product may still be used as a proxy through which we may be able to estimate diffuse attenuation (Figure 4.1, Pinkerton unpublished data).

Austin & Petzold (1981) give a relationship between chlorophyll concentration and the diffuse attenuation coefficient at 490 nm. Austin & Petzold (1986) give data to allow the diffuse attenuation coefficient at any wavelength between 350–700 nm to be estimated from that at 490 nm. For a given chlorophyll concentration, we used data in these references to model spectral downwelling irradiance at depth. Spectral irradiance values were converted to PAR at each depth using Equation [4.2]. The PAR attenuation coefficient was then estimated by least squares regression in log-light linear-depth space. This procedure was repeated for a variety of chlorophyll concentrations, to give the relationship between the downwelling (cosine) PAR attenuation coefficient and chlorophyll- concentration. A quartic regression in log-log space was fitted to the data using a least squares method. This relationship agrees very closely with measurements of the relationship between K_E and chlorophyll concentration in a variety of open ocean and coastal New Zealand waters (Figure 4.1). The validation shown in Figure 4.1 should be considered preliminary, as it includes measurements from only a limited number of New Zealand coastal regions. Further work on this issue is ongoing at NIWA.

For the purposes of this work, we assume that the underwater diffuse downwelling attenuation (K_E) is invariant with depth (i.e. we assume a well-mixed water column), and does not vary with time of day. The assumption that the inherent optical properties of the water column (and hence the diffuse attenuation coefficient) are invariant with depth is likely to hold where vertical mixing extends between the water surface and the sea-bed. This will include coastal waters more than ~1 km from shore where density-driven vertical stratification has broken down by wind mixing, and which do not have sub-surface chlorophyll maxima. These areas will include most of the waters for which we envisage the data being most useful within the Marine Environment Classification. For open ocean waters that have a permanent thermocline, sub-surface chlorophyll maximum, or waters very close to freshwater input (within ~1 km) where coloured material such as sediment may be held close to the surface by density-driven stratification, the assumption will be invalid. In these cases, information on the depth-distribution of diffuse attenuation coefficient is needed to improve our estimate of light at the sea-bed. The data required for this is not yet available for the New Zealand region.

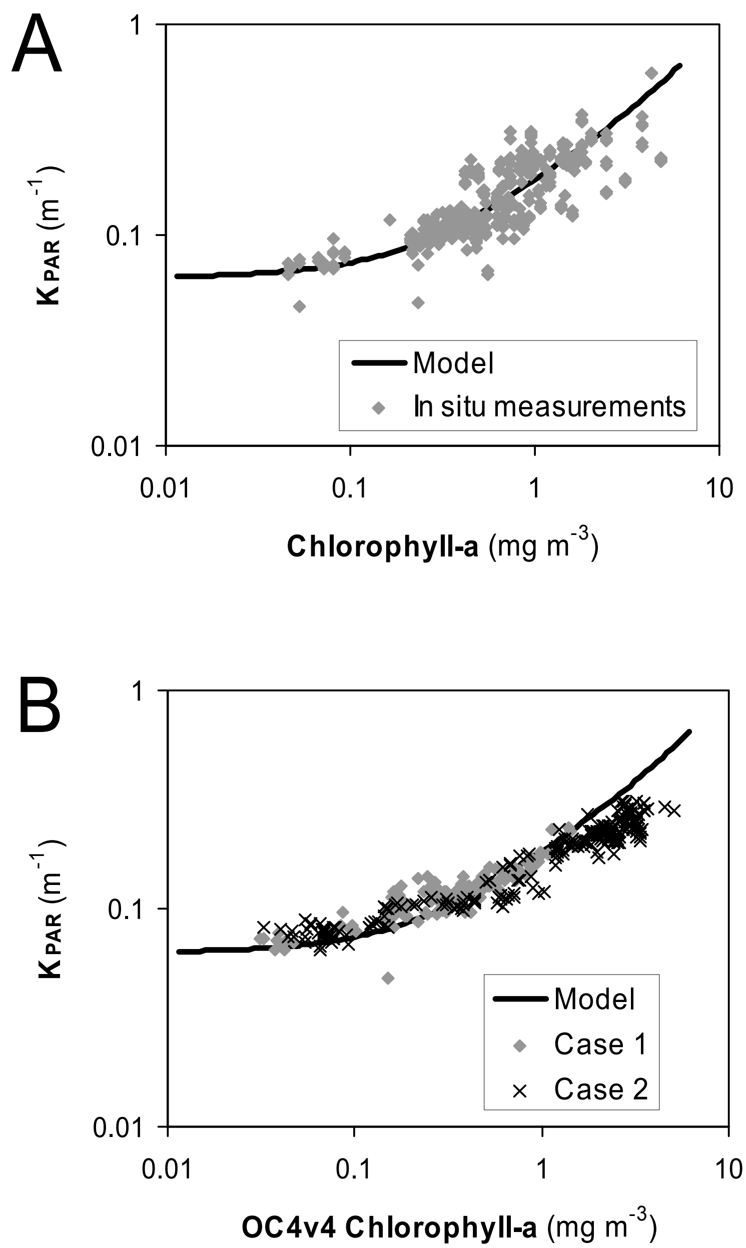

Figure 4.1. **Estimation of the downwelling diffuse attenuation coefficient for PAR (K_{PAR}=K_E) from chlorophyll concentration. In both panels, the black line is the modelled relationship used in this study, and derived as explained in the text. A: The grey symbols are in situ measurements of chlorophyll and PAR attenuation in a variety of New Zealand coastal and open-ocean waters. B: Chlorophyll concentration estimated by the OC4v4 algorithm. The grey squares indicate Case 1 conditions, and the crosses show Case 2 conditions. The model fits both water types reasonably well. Concentrations of chlorophyll estimated by the OC4v4**

algorithm less than 0.05 mg m^{-3} and greater than 3 mg m^{-3} make up a minor proportion of the data used in the production of the seabed PAR layer.

4.8 Results

Examples of incident irradiance, cloud cover and chlorophyll are given in Figures 4.2, 4.3 and 4.4. A given pixel was obscured by cloud on average for 83% of the time.

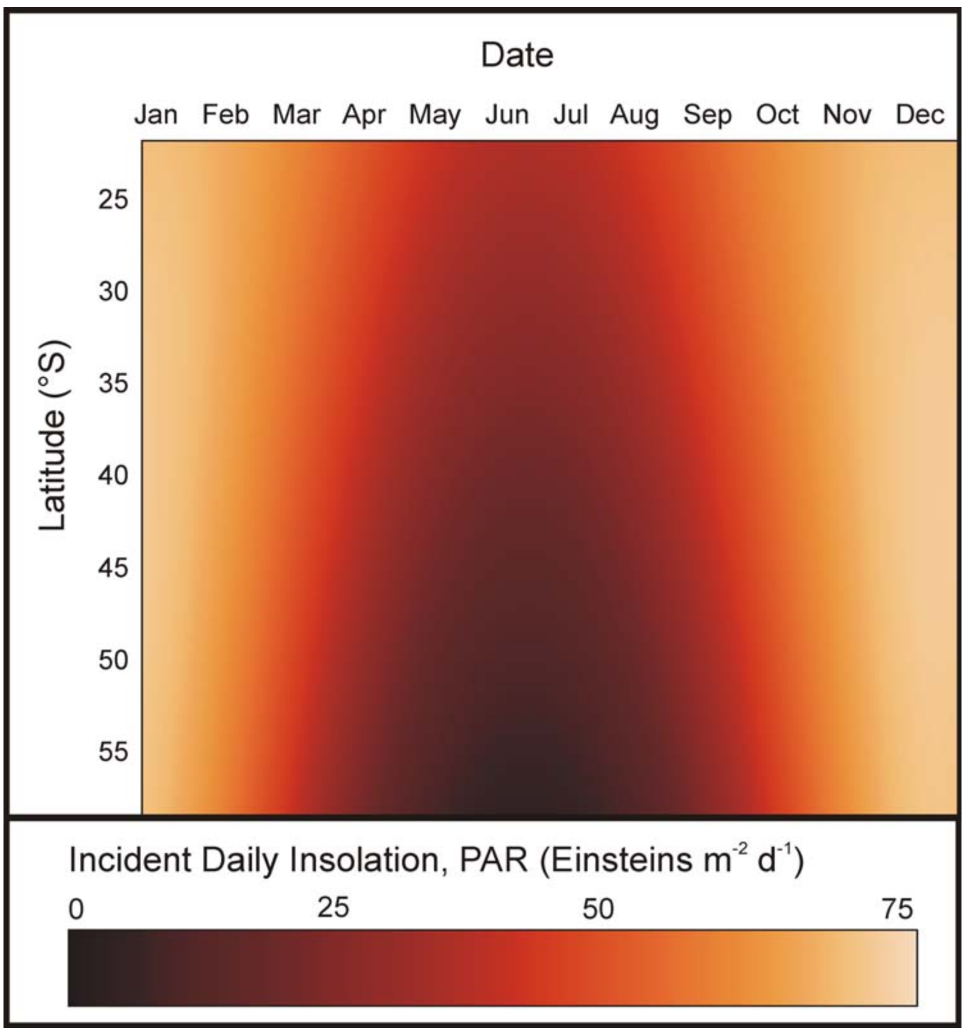

Figure 4.2. Daily incident insolation for the study region with latitude and day of year. The value plotted is equivalent to the factor $F_0.D.t.R$ in Equations [4.4] and [4.5]

Continue to next file: MCU6c