$$L_W(\lambda,t) = 0.543 L_u(0^-,\lambda,t). \tag{3.7}$$

Given only discrete radiometric measurements, $\bar{K}_L(z_i,\lambda)$ must be estimated as some combination of the methods 1a) through 1d), described above, for determining the diffuse attenuation coefficient. Some possible approaches are:

a. If upwelled radiance is measured using two wire-mounted radiometers at depths $z_2 > z_1$, for example, $K_L(\bar{z}_{12},\lambda,t)$ may be calculated with (3.5) for the interval between the two depths. [Combining radiance measurements in this way between hull-mounted and wire-mounted radiometers, *i.e.* to determine $K_L(\bar{z}_{01},\lambda,t)$, may be less straightforward if the buoy diameter is large and/or the measurement at z_O is not nadir-viewing] In clear oligotrophic water masses, it may be reasonable to assume optical homogeneity from the surface to depth z_2, or that $\bar{K}_L(z_1,\lambda,t) \cong K_L(\bar{z}_{12},\lambda,t)$. Other approximations must be considered if there is reason to believe that optical properties vary strongly in the layer above depth z_2.

b. If upwelled radiance is measured at only one depth, whether using a hull-mounted or wire-mounted radiometer, and downwelled irradiance is measured at one or more depths, it can usually be assumed that $K_L(z,\lambda) \approx K_d(z,\lambda)$ within approximately 5 % (Kirk 1994). Then, $\bar{K}_L(z_i,\lambda)$ may be determined using some combination of (3.3) and (3.4).

c. If upwelled radiance is measured only at depth z_o, just beneath the buoy hull, then there is no choice but to assume that $\bar{K}_L(z,\lambda) \approx \bar{K}(\lambda)$ and apply remote sensing algorithms such as those cited above under 1a).

3. ***Normalized Water-Leaving Radiance*** $L_{WN}(\lambda,t)$ is calculated from $L_W(\lambda,t)$ following the definition of (Gordon and Clark 1981)[7] as

$$L_{WN}(\lambda,t) = L_W(\lambda,t)\frac{\bar{F}_o(\lambda)}{E_S(\lambda,t)}, \tag{3.8}$$

where $\bar{F}_o(\lambda)$ is mean extraterrestrial solar irradiance (Neckle and Labs 1984) [see also Vol. I, Chapter 2, equation (2.55) in and Vol. III, Chapter 4 (Sect. 4.1)]. If reliable measurements of $E_S(\lambda,t)$ are available, they are substituted directly in (3.8). Otherwise, incident surface irradiance may be approximated either as the modeled clear-sky irradiance $\tilde{E}_S(\lambda,t)$ (*e.g.* Frouin *et al*. 1989, Gregg and Carder 1990) calculated for the solar zenith angle θ_o at time *t*, or more simply as

$$\tilde{E}_S(\lambda,t) = \bar{F}_o(\lambda)t_{atm}(\lambda,\theta_o)\cos\theta_o\left(\frac{d_o}{d}\right)^2, \tag{3.9}$$

where at time *t*, $t_{atm}(\lambda,\theta_o)$ is the diffuse transmission of the atmosphere, and d_o and *d* are the mean and actual earth-sun differences, respectively. Finally, $L_{WN}(\lambda,t)$ must be converted to Exact Normalized Water-Leaving Radiance by the methods described in Vol. III, Chapter 4.

4. ***Ocean Color Remote Sensing Parameters, Chl*** – chlorophyll a concentration in mg m^{-3} – and ***K490*** – the diffuse attenuation coefficient in m^{-1} averaged over the first e-folding attenuation depth - are

[7] Gordon *et al*. (1988) introduced a variant definition of Normalized Water-Leaving Radiance that included an adjustment for the downward Fresnel transmittance of incident direct solar flux into the ocean. Although this approximation has the correct sign, its magnitude is not correct for a real ocean surface (even under calm conditions). Therefore, the Gordon *et al*. (1988) definition is not used, because it is inconsistent with the definition of Exact Normalized Water-Leaving Radiance (Vol. III, Chapter 4), which correctly accounts for downward irradiance transmittance through the sea surface.