112

Estimation of irradiance just below the air-water interface

Heidi M. Dierssen and Raymond C. Smith

Institute for Computational Earth System Science and Department of Geography
University of California Santa Barbara
Santa Barbara, California 93106

ABSTRACT

Measuring irradiance just beneath the air-water interface, $E_d(0^-)$, is challenging because of environmental variability of the incident radiation field, such as effects of waves, perturbation by the instrument platform (ship shadow) and instrument limitations (i.e., size, orientation). Accurate measurements of subsurface irradiance and radiance, however, are critical in the estimation of remote-sensing reflectance values and the development of ocean color algorithms. Subsurface irradiance is typically estimated by extrapolating measured near-surface underwater spectra back to just beneath the surface. Such an approach, assumes that the water's optical properties are consistent within the extrapolation interval. However, the diffuse attenuation coefficients can vary widely in the surface layer due to selective absorption of the short and long wavelengths, pigment concentrations, and ship shadow effects and are strongly dependent on the sampling depth used in the calculation. Another independent estimate of $E_d(0^-)$ is derived by propagating irradiance measured above the sea surface to just beneath the air-water interface. Here, we compare these two independent estimates of $E_d(0^-)$ to examine the accuracy of our methods and instrumentation. We use measurements of downwelling spectral irradiance collected over two seasons at Palmer Station, Antarctica using a Profiling Reflectance Radiometer (PRR) deployed in freefall mode from a small zodiac, so as to minimize ship shadow effects. While estimates of $E_d(0^-)$ made from above and below the sea surface data were highly correlated for overcast days, clear days showed much more scatter between the two estimates. This was attributed to wave effects and the lack of completely clear skies without haze or high clouds. Comparison of above and below water observations with theoretical computations suggest systematic error in immersion coefficients used to calibrate the instrument. Further, very shallow (1-2m) density structure introduces layers of water with differing optical properties and causes error in the estimation of $E_d(0^-)$.

Keywords: ocean optics, irradiance, air-water interface, immersion coefficient, transmittance, diffuse attenuation coefficient

1. INTRODUCTION

Estimates of irradiance and radiance just beneath the air-water interface ($E_d(0^-)$ and $L_u(0^-)$) are used to develop and validate algorithms for remotely sensed ocean color observations. When propagated through the sea surface, $L_u(0^-,\lambda)$ is equivalent to the water-leaving radiance $L_w(0^-,\lambda)$ which, neglecting atmospheric effects, is the signal measured from a remote platform (e.g., satellites, planes). Further, $E_d(0^-,\lambda)$ can be used to normalize estimates of $L_w(0^-,\lambda)$ to account for the variability in solar zenith angle and atmospheric conditions. Measuring these quantities at an infinitesimal depth below the sea surface, however, is difficult due to instrument limitations (i.e., size, orientation, shadowing), perturbation by the instrument platform, fluctuations in cloud cover, and waves. In lieu of direct measurement, one technique has been to extrapolate from a measured profile of irradiance/radiance versus depth to estimate a value just beneath the sea surface. This approach assumes that the diffuse attenuation coefficients of upwelled and downwelled light, K_d and K_L, do not vary over the depth interval used in the extrapolation. However, estimates of K_d and K_L can vary widely within the surface layer due to selective absorption of the short and long wavelengths, pigment concentrations, and ship shadow effects. Consequently, extrapolation results can be strongly dependent on the sampling depth used in the regression calculation.

Another method of obtaining $E_d(0^-)$ is to propagate estimates of irradiance made above the sea surface, $E_d(0^+)$, through the air-water interface. These estimates of $E_d(0^-)$, hereafter referred to as $+E_d(0^-)$, should be comparable to the estimates made from below the sea-surface, hereafter referred to as $-E_d(0^-)$. Additionally, such estimates can be used to correct data for shadowing due to the ship's presence[1]. Ship shadow can be especially significant during overcast conditions[1] and when optical instruments are deployed less than 1 m off the stern of a ship[2]. This paper attempts to reconcile values of $+E_d(0^-)$ and $-E_d(0^-)$ which have been estimated from data which is not perturbed by ship and shadowing problems.

The polar environment can provide additional difficulty in obtaining accurate near-surface radiation estimates. The presence of ice near the instrumentation can contaminate irradiance measurements and lead to a sampling bias of just the open water. Further, melting of ice from glaciers and pack ice can lead to a surface lens of fresh water that can cause relatively

SPIE Vol. 2963 • 0277-786X/97/$10.00

large changes in the diffuse attenuation over just a few meters in water depth. At Palmer Station, extensive cloud cover serves to reduce the theoretical clear sky daily photosynthetically available radiation (PAR) on average by 40% in the spring time and 50% during the remainder of the year[3]. The extreme southern latitude of this site also means that the solar angle is always greater than 40 degrees and estimating the transmittance through the air-water interface more difficult.

2. METHODS

2.1 Data Collection

This paper utilizes two datasets collected in the 94/95 and 95/96 field seasons at Palmer Station (64°46S, 64°03W). Weather and ice permitting, data was collected weekly from a series of coastal stations during the season of maximum production (November to March). A Profiling Reflectance Radiometer (manufactured by Biospherical Inc. (BSI)) with wavebands centered at 412, 443, 490, 510, 555, and 665 (or 656) nm was deployed in freefall mode at a distance from the zodiac so as to minimize ship shadow effects[4]. Different PRR units with nearly identical instrument configuration were used in the 94/95 and 95/96 field seasons: the 94/95 PRR contained a channel centered at 656 nm and the 95/96 PRR contained a channel centered at 665 nm. Simultaneous measurements of downwelling irradiance were taken for these same wavebands from the deck of the zodiac. In order to completely avoid ship shadow perturbations of the light-field, SeaWiFS Ocean Optics Protocol[5] provide estimates of the minimum distance to deploy optical instrumentation away from the ship, expressed in terms of attenuation lengths in the water column. Based on the mean upwelling and downwelling attenuation coefficients measured during the 94/95 field season, the recommended average distance for deployment of the PRR for measuring downwelling irradiance is approximately 4 m and for upwelling radiance is 9 m. The PRR was typically deployed 15-20m from the zodiac.

2.1 Estimating $E_d(0^-)$ from above sea surface $(+E_d(0^-))$

Smith and Baker (1986)[1] show that to within a few percent, the downward irradiance just beneath the sea surface, $E_d(0^-)$, can be calculated from the irradiance measured just above the sea surface, $E_d(0^+)$ as follows:

$$E_d(0^-) \approx 1.03\, t(\lambda,\theta)\, E_d(0^+) \tag{1}$$

The global transmittance, $t(\lambda,\theta)$ includes reflectance of both direct (ρ_{sun}) and diffuse (ρ_{sky}) irradiance, such that:

$$t(\lambda,\theta) = (1-\rho_{sun}(\theta))\,(1-y(\lambda,\theta)) + (1-\rho_{sky}(\theta))\,y(\lambda,\theta) \tag{2}$$

In this formulation, the parameter y (or y-ratio) is the ratio of diffuse to total irradiance; ρ_{sun} is the Fresnel reflectance of direct radiation; and ρ_{sky} is assumed to be approximately 0.066. On a cloudy day, the y ratio approaches 1.0 and the above equation, assuming ρ_{sky} is 0.066 [1,6], reduces to:

$$E_d(0^-) \approx 0.96\, E_d(0^+) \tag{3}$$

For sunny days, the Fresnel reflectance was used to estimate ρ_{sun} in Equation 2. The y-ratio was assumed to be 0 for sunny days and the possible consequences of this decision are discussed in the results section.

2.2 Estimating $E_d(0^-)$ from below sea surface $(-E_d(0^-))$

Profiles of $E_d(z,\lambda)$were measured over the upper few optical depths and estimates of $E_d(0^-)$ were obtained by propagating $E_d(z,\lambda)$ back to the surface using a least squares regression technique [7,8]. Mathematically:

$$K_d = \frac{-d \ln E_d}{dz} \tag{4}$$

$$\ln E_d(0^-) = \ln E_d(z) + K_d z \tag{5}$$

For clear waters, using the same depth interval to obtain estimates of $E_d(0^-)$ is appropriate[8]. Choosing an appropriate averaging depth for the red region of the spectrum, which attenuates rapidly in the water column, can be problematic. For coastal waters with considerable vertical structure in the water column, moreover, difficulties can arise in choosing an averaging depth that is representative of the near-surface waters. We found that extrapolating irradiance over a specified optical depth was more appropriate both conceptually and experimentally.

3. RESULTS

Figures 1a and 1b are comparisons of $+E_d(0^-)$ and $-E_d(0^-)$ for overcast days from the 94/95 and 95/96 field seasons, respectively. The vertical axis contains estimates of $+E_d(0^-)$ from the deck unit (Eq. 3) and the horizontal axis contains $-E_d(0^-)$ estimated by extrapolating the measured in-water irradiance to just beneath the surface (Eqs. 4 and 5). An extrapolation depth corresponding to 1 optical depth was used. While there is a good correlation between these two methods (r^2=0.97-0.99), the 94/95 data do not have a 1:1 data. The estimates of $E_d(0^-)$ from measured in-water irradiance are higher for all channels than those derived from $E_d(0^+)$ and frequently from the measured $E_d(0^+)$ itself. While this could be possible for a few stations with high near-surface reflectance, this is highly unlikely for all sampling stations. The 95/96 data, however, is very close to a 1:1 correspondence for all channels. As expected, the lowest correlation (0.95) occurs for the red channel (656/665 nm) due to the rapid attenuation of these wavelengths by water itself. As noted earlier, two different PRR instruments were used for these two field seasons.

Figures 2a and 2b are clear sky comparisons of $+E_d(0^-)$ and $-E_d(0^-)$ for the 94/95 and 95/96 seasons. Reflectance has been calculated using Fresnel reflectance and the assumption of no cloud influences (y-ratio=0). Fewer datapoints are in this analysis as clear skies occur infrequently. As shown, both seasons show greater scatter in the data, with r^2 values as low as 0.6. Moreover, the 94/95 field season shows an even more exaggerated trend of higher $-E_d(0^-)$ to $+E_d(0^-)$ ratio than was found for cloudy skies and the 95/96 data no longer exhibits a 1:1 correspondence. One reason for the large amount of scatter in this data is due to the assumption that the y-ratio is 0 for clear skies. Figure 3 presents the results of modeling the direct and diffuse portions of the clear sky irradiance for different sampling days and times at Palmer Station using SBDart[9]. Because of the large atmospheric pathlength that photons travel in polar regions, the diffuse portion of the skylight, especially for the blue wavelengths, can be as high as 50% of the total irradiance. Depending on the day and time (i.e., ρ_{sun} varying from 0.02 to 0.07) and the wavelength, this would cause a few percent difference in the transmittance. If this were the only factor, however, the data from the blue wavelengths would be much more scattered than from the red wavelengths, which is not observed. Another reason for the observed scatter is the high variability in incoming irradiance during the casts. While the solar disc may be visible on these days, and hence by "Antarctic standards" are judged to be clear, they are, in fact, often hazy or contain high clouds that cause large variability in incoming insolation. Such atmospheric variability, especially given the larger solar angles in this region, make estimating $-E_d(0^-)$ relatively difficult. Variations in incoming insolation on clear days also indicate that the y-ratio should be greater than under clear skies. Because global transmittance is relatively insensitive to changes in the y-ratio for solar angles less than 70° [1], however, this would only change the value of $+E_d(0^-)$ by a few percent. Finally, wave slopes on the ocean surface can increase transmittance through the air-water interface, especially at high solar angles. At the solar angles corresponding to our sampling, this could cause a few percent difference in transmittance. There are too few days with completely clear skies and calm seas to statistically evaluate the impact of waves on the estimates of $+E_d(0^-)$. The cumulative impact of all of these factors contributed to the high scatter in the clear sky data.

The $-E_d(0^-)$ from above were calculated using an optical depth of 1. However, we also evaluated the impact of using various other optical depths to estimate $-E_d(0^-)$ in relation to estimates of $+E_d(0^-)$. Figure 4 presents the mean values of $-E_d(0^-)/+E_d(0^-)$ and one standard deviation from the mean for the 94/95 and 95/96 data. The wavebands centered at 412 nm and 656/665 nm are representative of the remaining bands. For 94/95, $-E_d(0^-)$ is about 8-10% higher than $+E_d(0^-)$ regardless of the optical depth used in the extrapolation. The 95/96 field data show less than 4% difference between $-E_d(0^-)$ and $+E_d(0^-)$ regardless of the optical depth used in the extrapolation. Further, the 94/95 errorbars are generally larger, especially with depth, than the 95/96 errorbars. This difference is primarily because several stations in 94/95 contained a surface lens of fresh water with different optical properties from the underlying water column. Figure 5a illustrates the physical and optical water column properties for one of these stations. As shown, the first few meters of water are fresher and less dense than the underlying water column, and the diffuse attenuation coefficient for PAR (K_{par}) drops off dramatically within this surface layer. Figure 5b shows a very strong bias in the depth interval chosen to estimate $-E_d(0^-)$ also for this station. The estimate of $-E_d(0^-)$ using 10 m was 50% of that estimated using 2 m. Water structure of this type is not as readily observed in data obtained from larger ships because of the mixing of surface waters caused by movement of the vessel.

4. DISCUSSION

Estimates of $E_d(0^-)$ making use of above-water and in-water profiles can be made independently of each other. Comparisons should be made of these independent estimates to provide a check of both the methodology and instrumentation used in sampling programs. We find that estimates of $E_d(0^-)$ from below and above the sea surface were highly correlated for both field seasons for the overcast days. For the 94/95 PRR, however, $-E_d(0^-)$ was consistently 8-10% higher than estimates of $+E_d(0^-)$. One reason for such a bias, could involve the methods used to estimate $E_d(0^-)$. As shown in Figure 3, no systematic bias was found by using different depth intervals in estimating $-E_d(0^-)$. Such a bias could occur if the transmittance through the air-water interface on overcast days was greater than the 0.96 used above. However, this

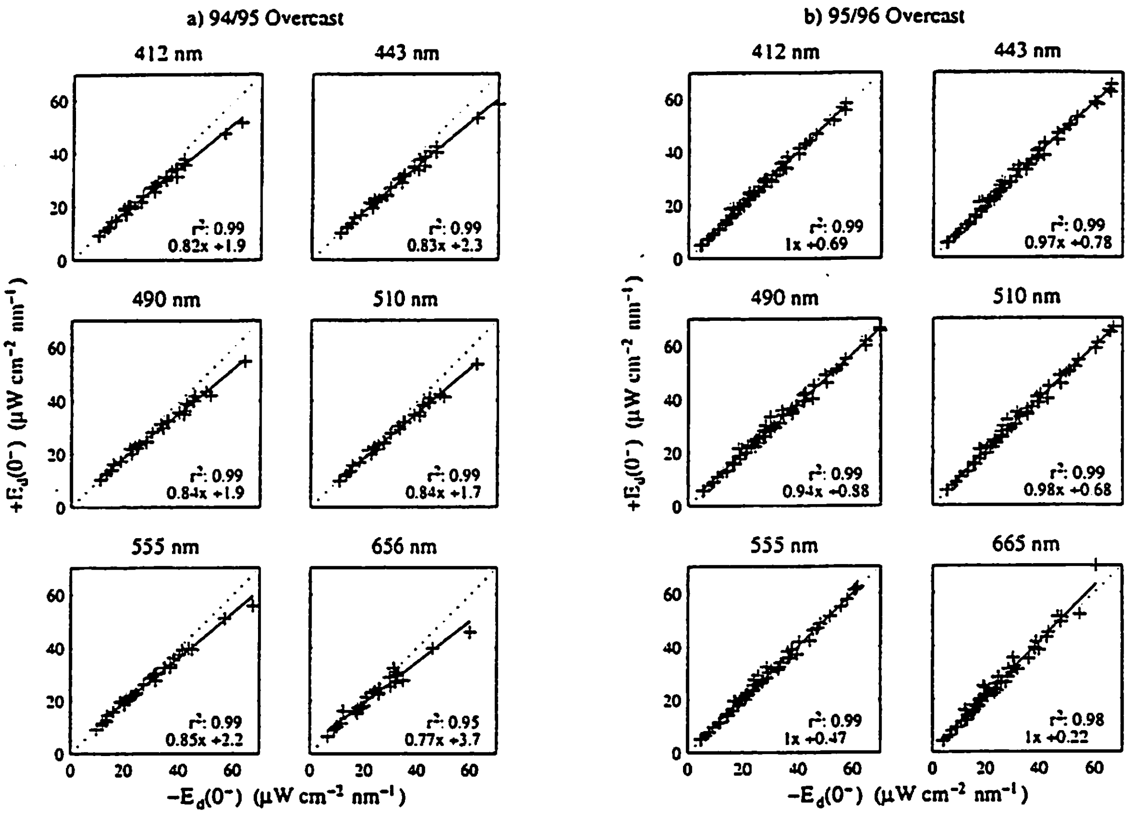


Fig. 1 Estimates of spectral irradiance just beneath the air-water interface made from above-water, $+E_d(0^-)$, and in-water, $-E_d(0^-)$, measurements. Data is from overcast days sampled around Palmer Station, Antarctica. (a) 94/95 field season. (b) 95/96 season, using a different PRR.

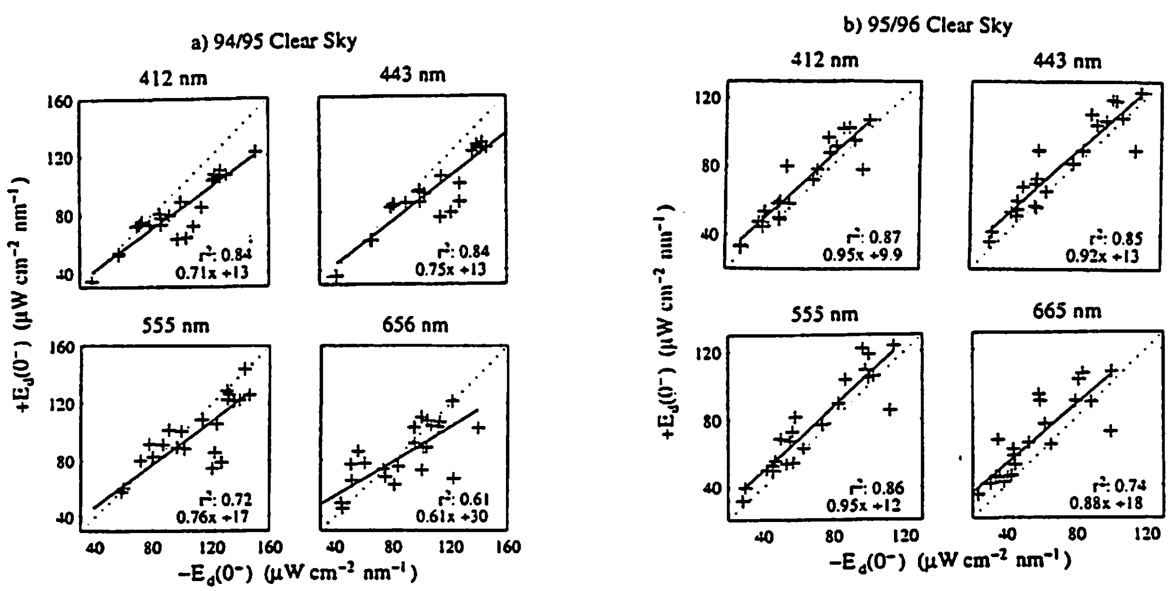


Fig. 2 Estimates of spectral irradiance just beneath the air-water interface made from above-water, $+E_d(0^-)$, and in-water, $-E_d(0^-)$, measurements. Data is from mostly clear days sampled around Palmer Station, Antarctica. (a) 94/95 field season. (b) 95/96 season, using a different PRR.

assumption appeared to be correct for the 95/96 data and even if transmittance is assumed to be 1.00 (highly unlikely), $-E_d(0^-)$ is still found to be 4-6% higher than estimates of $+E_d(0^-)$ for the 94/95 data. In addition, one would not expect to find systematic biases due to improper tilting of the instrumentation. While these factors could account for some scatter in the data, they generally do not create a consistent bias.

The most probable explanation for the systematic differences between the 94/95 instruments involves problems with the instrument calibration, particularly the immersion coefficients. Both the deck unit and the PRR unit are calibrated in air, but an immersion coefficient is applied to the PRR to account for differences in reflectance between the collector in air and in water[10]. Some transmission is lost when the collector is submerged in water. The same immersion factors (supplied by BSI) were applied to the calibrations of both the 94/95 and 95/96 PRR instruments. However, individual collectors, even with the same materials and design specifications, can have widely varying immersion coefficients. Mueller (1995)[11] measured the total range between immersion coefficients of 12 collectors to be as high as 15% at some wavelengths. If immersion coefficients are decreased by 8-10% for the 94/95 PRR, the two estimates of $E_d(0^-)$ correspond to within a few percent. As estimates of $-E_d(0^-)$ and $+E_d(0^-)$ are on average within a few percent of each other for the 95/96 data, the immersion coefficients for the 95/96 PRR appear to be nearly correct. An analysis of the immersion coefficient for both instruments is in progress.

In conclusion, comparisons of the two independent estimates of $E_d(0^-)$, one from above and one from below the sea surface, can provide useful information about the water properties, air-sea interactions, and the instrumentation and methods used to obtain the data. The very high correlation between $-E_d(0^-)$ and $+E_d(0^-)$ for overcast days suggests that the methods used to make these estimates are consistent. However, the systematic offsets between these estimates of $E_d(0^-)$ for the 94/95 data suggest systematic error in the immersion coefficients used in the calibration of the PRR. For clear skies, the lower correlation between these two estimates of $E_d(0^-)$ is likely due to a combination of physical effects: the high fraction of diffuse sky in polar regions on clear days, the presence of fog, haze, and clouds during "clear" days that cause a high level of variability in incoming radiation during each cast; and the presence of waves which change the reflectance of the direct solar beam used to estimate $+E_d(0^-)$. As remotely sensed ocean color observations will only be possible on clear days, developing algorithms and interpreting remotely sensed data from this region will be challenging. In addition, large biases in estimates

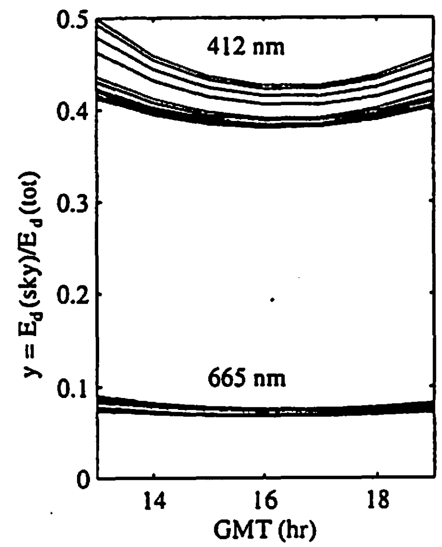


Fig. 3 Estimates of the y-ratio for clear skies at Palmer Station for different sampling days and times. Estimates made using SBDart with a 10 nm spectral band-width centered at both 412 nm and 665 nm.

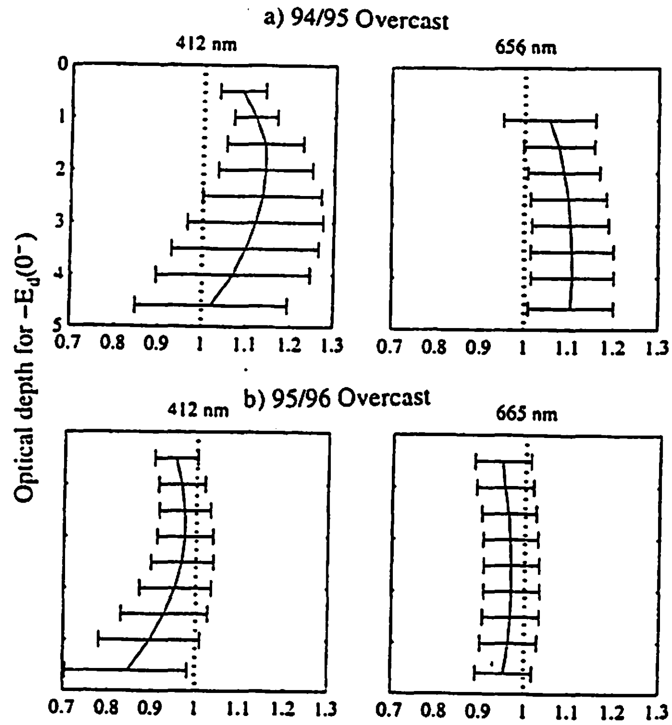


Fig. 4 Mean $+E_d(0^-)/-E_d(0^-)$ for all stations. Errorbars are one standard deviation from the mean. The vertical axis is the depth interval, in optical depth, used to estimate $-E_d(0^-)$. (a) 94/95 season. (b) 95/96 season.

of $-E_d(0^-)$ can arise when ice melt, unique to polar regions, produces a shallow (1-2 m) surface layer of fresh water with differing optical properties from the underlying water column. Such conditions are not typically encountered with sampling from larger ships that cause mixing of the surface layer of water and are better dealt with using optical depths, rather than discrete depths, to estimate $-E_d(0^-)$. Future refinement of this analysis would involve using a radiative transfer model to estimate $E_d(0^-)$ and comparing the modeled estimates to those calculated here.

5. ACKNOWLEDGMENTS

This research was funded by the Palmer LTER, National Science Foundation Office of Polar Programs (DPP90-11927, to RCS), the National Aeronautics and Space Administration (NAGW 290-3. to RCS), and a California Space Grant Fellowship (to HMD). The authors would like to acknowledge David Menzies for sharing his time and expertise.

6. REFERENCES

1. Smith, R.C. & Baker, K.S.,"Analysis of ocean optical data II," *SPIE - Ocean Optics VIII* **637**,95-107 (1986).
2. Weir, C.T., Siegle, D.A., Michaels, A.F. & Menzies, D.W.,"In situ evaluation of a ship's shadow," *SPIE - Ocean Optics XII* **2258**,815-821 (1994).
3. Dierssen, H.M., Smith, R.C. & Vernet, M. in *AGU/ASLO, Ocean Sciences Meeting 12-16 Feb* San Diego, 1996).
4. Waters, K.J., Smith, R.C. & Lewis, M.R.,"Avoiding ship-induced light-field perturbation in the determination of oceanic optical properties," *Oceanography* **November**,18-21 (1990).
5. Mueller, J.L. & Austin, R.W., *Volume 5, Ocean Optics Protocols for SeaWiFS Validation* NASA Technical Memorandum 104566: Sea WiFS Technical Report Series, Goddard Space Flight Center, 1992.
6. Preisendorfer, R.W., *Hydrologic Optics* , U.S. Dept. of Commerce, National Oceanic and Atmospheric Administration, Honolulu, Hawaii, 1976.
7. Smith, R.C. & Baker, K.S.,"The analysis of ocean optical data," *SPIE - Ocean Optics VII* **489**,119-126 (1984).
8. Siegel, D.A., O'Brien, M.C., Sorensen, J., Konnoff, D.A. & Fields, E., *BBOP data processinng and sampling procedures* Institute for Computational Earth System Science, Santa Barbara, CA, 1995.
9. Ricchiazzi, P., Gautier, C. & Lubin, D.,"Cloud scattering optical depth and local surface albedo in the Antarctic: Simultaneous retrieval using ground-based radiometry," *Journal of Geophysical Research* **100**(D10), 21091-21104 (1995).
10. Smith, R.C.,"An underwater spectral irradiance collector," *Journal of Marine Research* **27**(3), 341-351 (1969).
11. Mueller, J.L. in *Volume 27, Case Studies for SeaWiFS Calibration and Validation, Part 3.* (eds. Hooker, S.B., Firestone, E.R. & Acker, J.G.) (NASA Technical Memorandum 104566: Sea WiFS Technical Report Series, 1995).

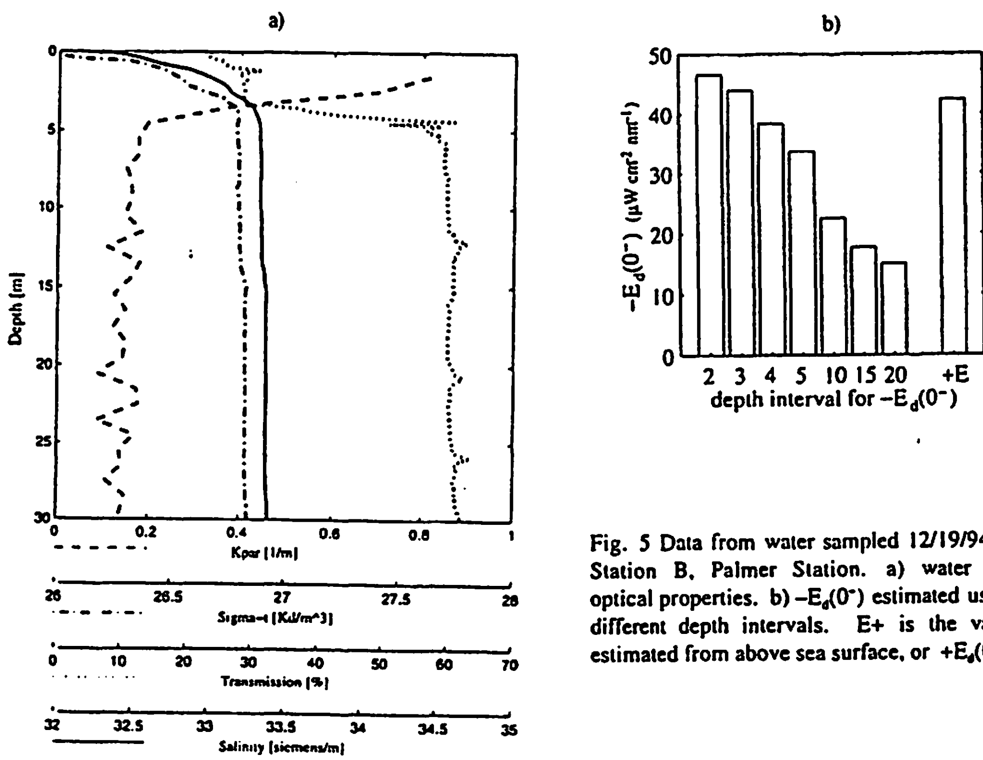
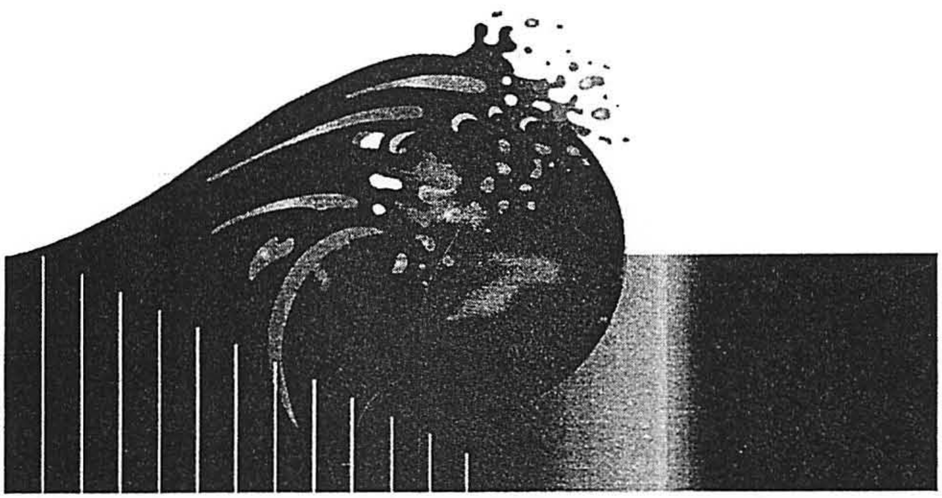


Fig. 5 Data from water sampled 12/19/94 at Station B, Palmer Station. a) water and optical properties. b) $-E_d(0^-)$ estimated using different depth intervals. E+ is the value estimated from above sea surface, or $+E_d(0^-)$.



OCEAN OPTICS XIII

Steven G. Ackleson
Robert Frouin
Convenors/Editors

22–25 October 1996
Halifax, Nova Scotia, Canada

Sponsors
Office of Naval Research (USA)
NASA (USA)
Russian Academy of Sciences
SPIE—The International Society for Optical Engineering

Associate Sponsors
American Geophysical Union
The Oceanography Society (USA)
American Society of Limnology & Oceanography
Bedford Institution of Oceanography (Canada)
Defense Research Establishment Valcartier (Canada)
Alliance for Marine Remote Sensing (Canada)

SPIE Volume 2963