detected narrowband signals with the expected frequency variation 40 dB below noise (in a 1-kHz band), while rejecting fixed-frequency (interference-like) signals at, or even above, the level of the noise.

Observations were made on target stars taken from the Royal Greenwich Observatory (RGO) catalog, which includes all known stars within 25 parsecs (82 light years) (*9*); all F, G, and K stars of luminosity class V (dwarf) or of unknown luminosity class from the RGO list that are not known to be members of multiple star systems and that can be seen with the Arecibo dish (0° < declination < 38°) were observed during this search. In addition, the flat-spectrum radio sources 4C05.64 and PKS0735+17 and the unusual pulsars 1913+16 (binary) and 1952+29 (extremely small rate of change of period) were observed. Altogether 185 objects were examined, 60 of them on more than one occasion, during 80 hours of telescope time. Ten 1-minute observations were made in the vicinity of each object, eight at the source position and two displaced from it by ±2 minutes in right ascension. In no case were extrastatistical narrowband signals detected. Terrestrial interference was not a problem, even during noise-prone daytime observations, confirming the reasoning presented earlier. There were no false alarms and hence no loss of observing time as a consequence of man-made interference. It is a simple matter to calculate the sensitivity of the overall system to narrowband polarized signals. Using the known parameters of the system, one obtains a sensitivity of 4×10^{-27} W/m^2 for a signal-to-noise ratio of 5:1 in each of the eight observations of each source; this is at least two orders of magnitude more sensitive than any previously reported SETI activity and corresponds to a total power incident on the earth's disk of less than a millionth of a microwatt. This can be expressed alternatively as the range at which this system could detect an identical antenna transmitting a 1-MW carrier, say, beamed at our sun. That distance is 370 parsecs, for a signal-to-noise ratio of 5:1 in each 1-minute observation.

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Decomposition of Calcium Carbonate and Organic Carbon in the Deep Oceans

Abstract. *Simple mass-balance calculations indicate that in seawater the calcium variation* (ΔCa) *correlates with the variation in the titration alkalinity* (ΔTA) *and the variation in the total carbon dioxide* ($\Delta\Sigma CO_2$) *or nitrate* (ΔNO_3) *according to the equations* $\Delta Ca = 0.46288\ \Delta TA + 0.074236\ \Delta\Sigma CO_2$ *and* $\Delta Ca = 0.5\ \Delta TA + 0.53125\ \Delta NO_3$. *The estimated values for* ΔCa *from these equations, which agree with the values obtained from direct measurements, have been used to estimate the ratio of the in situ inorganic to organic carbon fluxes in the oceans. The precise vertical distribution of this ratio is shown for the first time.*

The processes related to the abyssal dissolution of $CaCO_3$ and organic C provide a direct mechanism for the renewal of elements in the sea. They are also responsible for changing, or in fact maintaining in the steady state, the *p*H and buffer capacity of the oceans, and are of fundamental interest in chemical, biological, and geochemical oceanography (*1*). Values for $CaCO_3$ fluxes have been obtained from measurements of dissolved Ca in seawater (*2*). The analysis of dissolved Ca, in the presence of Mg and Sr, to a precision of ± 0.1 percent, is one of the classic problems in the analytical chemistry of seawater. Furthermore, the high concentration of dissolved Ca in seawater (10.3 mmole/kg for normal seawater at a salinity of 35 per mil) relative to the small changes due to $CaCO_3$ dissolution ($\Delta Ca < 0.1$ mmole/kg) makes the evaluation of Ca fluxes extremely difficult. Nonetheless, several recent papers have reported such analyses [see (*2–4*)].

The dissolution of $CaCO_3$ alone increases the Ca concentration and titration alkalinity (*TA*) in a fixed ratio of 0.5. Several investigators (*2, 4*) have therefore tried to linearly correlate observed Ca concentrations with *TA*; their success has been limited. The ratio is not exactly equal to 0.5 because the decomposition of organic material also changes *TA* but not Ca. Consequently, the ratio of Ca to *TA* is not a constant and one should not expect to find a linear correlation between Ca and *TA*. I examine here some recent chemical data on seawater in an attempt to correlate the Ca variation (ΔCa) with the variation in titration alkalinity (ΔTA) and in total CO_2 ($\Delta\Sigma CO_2$) or nitrate (ΔNO_3), and to estimate ΔCa from the available chemical data.

In seawater, $CaCO_3$ decomposes to form

$$CaCO_3 = Ca^{2+} + CO_3^{2-} \qquad (1)$$

The increases in Ca, *TA*, and ΣCO_2 are 1 mole, 2 equivalents, and 1 mole, respectively, for the dissolution of 1 mole of $CaCO_3$. The decomposition of 106 moles of organic C yields 106 moles of CO_2, 16 moles of HNO_3, and 1 mole of H_3PO_4 according to the Redfield-Ketchum-Richards model (*5*):

$$(CH_2O)_{106}(NH_3)_{16}H_3PO_4 + 138\ O_2 = 106\ CO_2 + 16\ HNO_3 + H_3PO_4 + 122\ H_2O \qquad (2)$$

In seawater, CO_2, HNO_3, and H_3PO_4 produced in situ react with CO_3^{2-} (*1, 6*) to form, respectively, HCO_3^-, NO_3^-, and HPO_4^{2-}, and Eq. 2 becomes

$$(CH_2O)_{106}(NH_3)_{16}H_3PO_4 + 138\ O_2 + 124\ CO_3^{2-} = 16\ NO_3^- + HPO_4^{2-} + 230\ HCO_3^- + 16\ H_2O \qquad (3)$$

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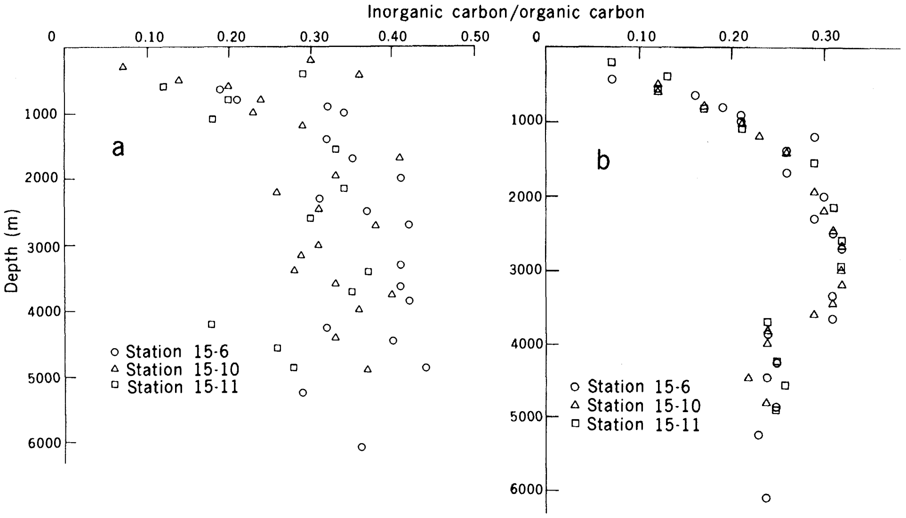

Fig. 1. The ratio of inorganic C to organic C flux versus the depth calculated from the ΔCa and ΔNO_3 data: (a) based on the use of measured ΔCa values and (b) based on the use of estimated ΔCa data.

The result from Eq. 3 is an increase of 106 moles in ΣCO_2 and a decrease of 17 equivalents in TA. The combined effect of decomposing x μmole of $CaCO_3$ and y μmole of organic matter in 1 kg of seawater on ΔTA (in microequivalents), $\Delta\Sigma CO_2$, (in micromoles), and ΔNO_3 (in micromoles) can be represented as follows:

$$\Delta TA = 2x - 17y \tag{4}$$

$$\Delta\Sigma CO_2 = x + 106y \tag{5}$$

$$\Delta NO_3 = 16y \tag{6}$$

Solving for x using Eqs. 4 and 5 and Eqs. 4 and 6 gives

$$\Delta Ca = x = 0.46288\ \Delta TA + 0.074236\ \Delta\Sigma CO_2 \tag{7}$$

or

$$\Delta Ca = x = 0.5\ \Delta TA + 0.53125\ \Delta NO_3 \tag{8}$$

Equation 8 is preferred because of the more general availability and the relatively higher precision of ΔNO_3 data by comparison with $\Delta\Sigma CO_2$ data.

Earlier workers (*2*, *4*) have tried to correlate ΔCa with ΔTA (or Ca with TA) by

$$\Delta Ca = \text{constant} + b\ \Delta TA \tag{9}$$

and have tried to evaluate the so-called theoretical slope, b. Equations 7 and 8 show that ΔCa is not a function of ΔTA alone and can vary (with $\Delta\Sigma CO_2$ or ΔNO_3), even with a constant ΔTA. I have used Eq. 8 to estimate ΔCa, using the chemical data taken from the Antipode 15 Expedition (Geochemical Ocean Section Study Intercalibration Experiment) Data Report (*7*). The values of ΔCa (measured), ΔTA, and ΔNO_3 have been calculated relative to a local surface seawater having a salinity, S, of 35.080 per mil, a Ca concentration of 10,332 μmole/kg, a TA of 2307 μeq/kg and a NO_3^- concentration of 0 μmole/kg. The concentration differences (ΔA) of property A are given by

$$\Delta A = A_{\text{in situ}} - A_{\text{ref}} \times \frac{S_{\text{in situ}}}{35.080} \tag{10}$$

The estimated ΔCa, which has an uncertainty of approximately ± 4 μmole/kg due to the uncertainties in ΔNO_3 (± 0.3 μmole/kg) and ΔTA (± 5 μeq/kg), agrees with the measured values, on the average, to ± 14.6 μmole/kg. This agreement is comparable to the precision of the direct measurements, ± 10 μmole/kg (*4*). The effect of water mixing and the contribution of other acids to the chemical processes have been neglected and may have slightly affected the validity of Eq. 4. For instance, if all the S in the organic material were oxidized to SO_4^{2-} (*1*, *8*), Eq. 4 would become

$$\Delta TA = 2x - 20.2y \tag{11}$$

The consequence is an increase of approximately 3.5 μmole/kg in the estimated ΔCa. This would bring the estimated ΔCa to a better agreement with the measured values.

The ratio of the in situ inorganic C to organic C variation (derived from the dissolution of $CaCO_3$ and organic material, respectively) can be calculated from the ΔCa and ΔNO_3 data by means of

$$\frac{\text{Inorg C}}{\text{Org C}} = \frac{x}{106y} = \frac{16\ \Delta Ca}{106\ \Delta NO_3} \tag{12}$$

The result based on the use of Eq. 12 and the measured ΔCa and ΔNO_3 data shows considerable scatter (Fig. 1a), which probably is due to the relatively large errors introduced in the measurement of the Ca concentration in seawater. Equation 12 has also been used together with the ΔNO_3 and estimated ΔCa data to calculate the same ratio. The result (Fig. 1b) indicates a gradual increase of the inorganic C contribution at greater depths. The contribution of the inorganic C reaches a maximum at 2800 m and tends to decrease below this depth. An abrupt decrease occurs at approximately 3650 m, which corresponds to the depth of the benthic front (*9*), and indicates that the bottom water comes from the Southern Ocean, which has a relatively low inorganic C flux.

The equations given above represent only the first approximations of the very complicated system in the oceans. The water masses have been assumed to come from the same surface origin, and the effects of mixing have been neglected. Nevertheless, this method is simple and the result reveals details that have not been shown before. Overall, the $CaCO_3$ contributes approximately 20 percent of the total C input (taking 0.25 as the average ratio of inorganic C to organic C) in the water column in the South Pacific Ocean. This value agrees with the work of Ben-Yaakov (*10*), is lower than the value of Li *et al.* (*11*) and Craig (*12*), but is higher than the data of Edmond (*13*).

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