Table 2 Countries operating 10 or more oceanographic research ships

Russia	86
United States	84
Japan	66
China	17
Ukraine	14
Korea	11
Germany	11
United Kingdom	10
Canada	10
All others (39)	111
Total	420

marine sciences they support has been the International Ship Operators Meeting, an intergovernmental association, founded in 1986, comprising representatives from various ship operating agencies that meets periodically to exchange information on ship operations and schedules, and work on common problems affecting research vessels. In 1999, 21 ship-operating nations were represented and extended membership by other states is ongoing.

Future Oceanographic Ships

The interest in, and growth of, marine science over the past half century shows little or no indication of diminishing. The trend in oceanographic investigations has been to carry larger and more complex instrumentation to sea. The size and capability of research vessels in support of developing projects has also increased.

Future oceanographic research ships can be expected to become somewhat larger than their counterparts today. This will result from demands for more sizeable scientific complements and laboratory spaces. Workdeck and shops will be needed for larger equipment systems such as buoy arrays, bottom stations, towed and autonomous vehicles. Larger overside handling systems incorporating motion compensation will make demands for more deck space.

There will be fewer differences between basic science, fisheries, and hydrographic surveying vessels so that one vessel can serve several purposes. This may result in fewer vessels, but overall tonnage and capacity can be expected to increase.

New types of craft may take a place alongside conventional ships. These include submarines, 'flip'-type vessels which transit horizontally and flip vertically on station, and small waterplane-area twin hull ships (SWATH). SWATH, or semisubmerged ships, are a relatively recent development in ship design. Although patents employing this concept show up in 1905, 1932, and 1946, it was not until 1972 that the US Navy built a 28 m, 220 ton prototype model. The principle of a SWATH ship is that submerged hulls do not follow surface wave motion, and thin struts supporting an above water platform which have a small cross-section (waterplane) are nearly transparent to surface waves, and have longer natural periods and reduced buoyancy force changes than a conventional hull. The result is that SWATH ships, both in theory and performance, demonstrate a remarkably stable environment and platform configuration which is highly attractive for science and engineering operations at sea.

See also

Coastal Zone Management. Fishing Methods and Fishing Fleets. International Organizations. Law of the Sea. Maritime Archaeology. Shipping and Ports.

SILICA

See **MARINE SILICA CYCLE**

SINGLE COMPOUND RADIOCARBON MEASUREMENTS

T. I. Eglinton and A. Pearson, Woods Hole Oceanographic Institution, Woods Hole, MA, USA

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Introduction

Many areas of scientific research use radiocarbon (carbon-14, ^{14}C) measurements to determine the age of carbon-containing materials. Radiocarbon's ~5700-year half-life means that this naturally

occurring radioisotope can provide information over decadal to millennial timescales. Radiocarbon is uniquely suited to biogeochemical studies, where much research is focused on carbon cycling at various spatial and temporal scales. In oceanography, investigators use the ^{14}C concentration of dissolved inorganic carbon (DIC) to monitor the movement of water masses throughout the global ocean. In marine sediment geochemistry, a major application is the dating of total organic carbon (TOC) in order to calculate sediment accumulation rates. Such chronologies frequently rely on the premise that most of the TOC derives from marine biomass production in the overlying water column.

However, the ^{14}C content of TOC in sediments, as well as other organic pools in the ocean (dissolved and particulate organic matter in the water column) often does not reflect a single input source. Multiple components with different respective ages can contribute to these pools and can be deposited concurrently in marine sediments (**Figure 1**). This is particularly true on the continental margins, where fresh vascular plant debris, soil organic matter, and fossil carbon eroded from sedimentary rocks can contribute a significant or even the dominant fraction of the TOC. This material dilutes the marine input and obscures the true age of the sediment. Although such contributions from multiple organic carbon sources can complicate the development of TOC-based sediment chronologies, these sediment records hold much important information concerning the cycling of organic carbon both within and between terrestrial and marine systems. The challenge, then, is to decipher these different inputs by resolving them into their individual parts.

Most of the allochthonous, or foreign, sources represent carbon with lower ^{14}C concentrations ('older' radiocarbon ages) than the fraction of TOC originating from phytoplanktonic production. The only exception is the rapid transport and sedimentation of recently synthesized terrestrial plant material, which is in equilibrium with the ^{14}C concentration of atmospheric CO_2. Other sources of nonmarine carbon typically are of intermediate (10^3–10^4 years) or 'infinite' (beyond the detection limit of 50–60 000 years) radiocarbon age, depending on the amount of time spent in other reservoirs such as soils, fluvial deposits, or carbon-rich rocks.

It is only at the molecular level that the full extent of this isotopic heterogeneity resulting from these

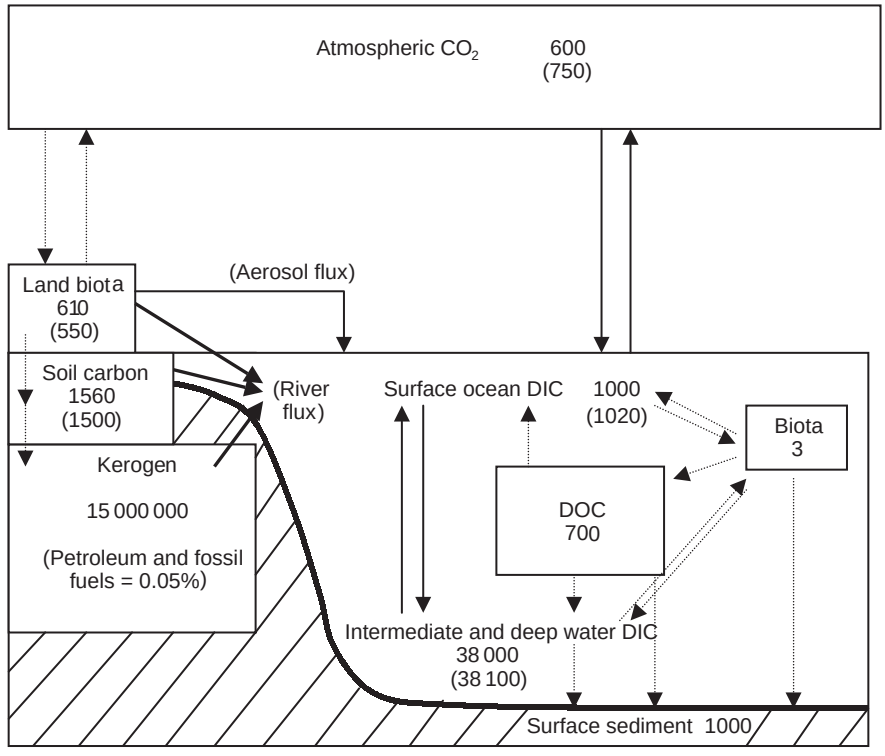

Figure 1 Major global reservoirs involved in active production, exchange and cycling of organic carbon. Reservoir sizes are shown in Gt carbon (1 GtC = 10^{15} g C). Numbers in parentheses are based on 1980s values; numbers without parentheses are estimates of the pre-anthropogenic values. Fluxes primarily mediated by biological reactions are shown with dashed arrows; physical transport processes are shown with solid arrows. (Modified after Siegenthaler and Sarmiento (1993) and Hedges and Oades (1997).)

diverse organic carbon inputs is expressed. Isotopic analysis of individual biomarker compounds was employed originally to study the stable carbon isotope (^{13}C) distribution in lipids of geological samples. It proved to be a useful tool to describe the diversity of carbon sources and metabolic pathways as well as to link specific compounds with their biological origins. Recently, this approach was expanded into a second isotopic dimension by the development of a practical method to achieve compound-specific ^{14}C analysis. Not only do these new ^{14}C analyses of individual biomarker molecules provide a tool for dating sediments, but they are another source of fundamental information about biogeochemical processes in the marine environment.

Carbon Isotopes

Carbon in the geosphere is composed of the stable isotopes ^{12}C (98.9%) and ^{13}C (1.1%), and the cosmogenic radionucleotide, ^{14}C (radiocarbon). Upon production, ^{14}C is incorporated quickly into atmospheric CO_2, where it occurs as approximately 10^{-10}% of the total atmospheric abundance of CO_2. The distribution of the minor isotopes relative to ^{12}C is governed by thermodynamic and kinetic fractionation processes[1], in addition to the radioactive decay associated with ^{14}C.

^{14}C Systematics

Today, most radiocarbon data are obtained through the use of accelerator mass spectrometry (AMS) rather than by counting individual decay events. In particular, the advantage of AMS is its small carbon requirement (micrograms to milligrams); this ability to analyze small samples is critical to the compound-specific ^{14}C approach, where sample sizes typically range from tens to hundreds of micrograms. These sample sizes are dictated by natural concentrations of the analytes in geochemical samples (often $< 1\,\mu g\,g^{-1}$ dry sediment), and by the capacity of the techniques used to isolate the individual compounds in high purity.

Raw AMS data are reported initially as fraction modern (f_m) carbon (eqn [1]).

$$f_m = \frac{R^{14/12}_{sn}}{R^{14/12}_{std}} \qquad [1]$$

$R^{14/12} \equiv {}^{14}C/{}^{12}C$ (some laboratories use $R = {}^{14}C/{}^{13}C$), sn indicates the sample has been normalized to a constant ^{13}C fractionation equivalent to $\delta^{13}C = -25$ppt, and std is the oxalic acid I (HOxI) or II (HOxII) modern-age standard, again normalized with respect to ^{13}C.

For geochemical applications, data often are reported as $\Delta^{14}C$ values (eqn [2]).

$$\Delta^{14}C = \left[f_m \left(\frac{e^{\lambda(y-x)}}{e^{\lambda(y-1950)}} \right) - 1 \right] \times 1000 \qquad [2]$$

Here, $\lambda = 1/8267(y^{-1})(= t_{1/2}/\ln 2)$, y equals the year of measurement, and x equals the year of sample formation or deposition (applied only when known by independent dating methods, for example, by the use of ^{210}Pb). This equation standardizes all $\Delta^{14}C$ values relative to the year AD 1950. In oceanography, $\Delta^{14}C$ is a convenient parameter because it is linear and can be used in isotopic mass balance calculations of the type shown in eqn [3].

$$\Delta^{14}C_{total} = \sum_i (\chi_i \Delta^{14}C_i) \qquad \sum_i \chi_i = 1 \qquad [3]$$

The 'radiocarbon age' of a sample is defined strictly as the age calculated using the Libby half-life of 5568 years (eqn [4]).

$$\text{Age} = -8033\, ln(f_m) \qquad [4]$$

For applications in which a calendar date is required, the calculated ages subsequently are converted using calibration curves that account for past natural variations in the rate of formation of ^{14}C. However, the true half-life of ^{14}C is 5730 years, and this true value should be used when making decay-related corrections in geochemical systems.

^{14}C Distribution in the Geosphere

Natural Processes

Atmospheric $^{14}CO_2$ is distributed rapidly throughout the terrestrial biosphere, and living plants and their heterotrophic consumers (animals) are in equilibrium with the $\Delta^{14}C$ value of the atmosphere. Thus, in radiocarbon dating, the ^{14}C concentration of a sample is strictly an indicator of the amount of time that has passed since the death of the terrestrial primary producer. When an organism assimilates a fraction of pre-aged carbon, an appropriate 'reservoir age' must be subtracted to correct for the deviation of this material from the age of the atmosphere. Therefore, reservoir time must be considered

[1] This article assumes the reader is familiar with the conventions used for reporting stable carbon isotopic ratios, i.e., $\delta^{13}C$ (ppt) $= 1000[(R/R_{PDB}) - 1]$ where $R \equiv {}^{13}C/{}^{12}C$. For further explanation, see the additional readings listed at the end of this article.

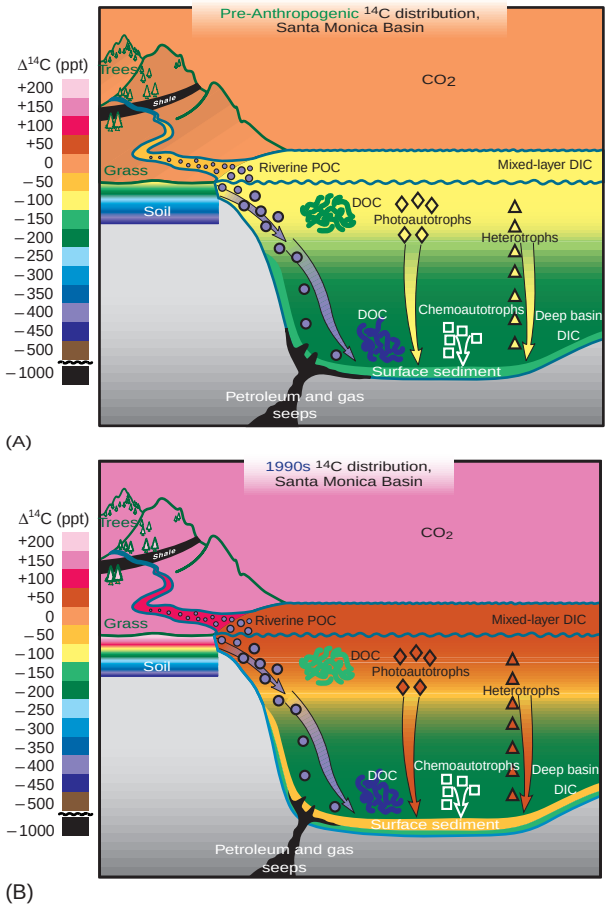

Figure 2 $\Delta^{14}C$ values for bulk carbon reservoirs in the region of Santa Monica Basin, California, USA: (A) prior to human influence ('pre-bomb'), and (B) contemporary values ('post-bomb'). Modified after Pearson (2000).)

when interpreting the ^{14}C 'ages' of all of the global organic carbon pools other than the land biota.

For example, continuous vertical mixing of the ocean provides the surface waters with some abyssal DIC that has been removed from contact with atmospheric CO_2 for up to 1500 years. This process gives the ocean an average surface water reservoir age of about 400 years ($\Delta^{14}C = -50$ ppt). A constant correction factor of 400 years often is subtracted from the radiocarbon dates of marine materials (both organic and inorganic). There are regional differences, however, and in upwelling areas the true deviation can approach 1300 years.

An example of the actual range of $\Delta^{14}C$ values found in the natural environment is shown in **Figure 2A**. This figure shows the distribution of ^{14}C in and around Santa Monica Basin, California, USA, prior to significant human influence. The basin sediments are the final burial location for organic matter derived from many of these sources, and the TOC $\Delta^{14}C$ value of -160 ppt represents a weighted average of the total organic carbon flux to the sediment surface.

Anthropogenic Perturbation

In addition to natural variations in atmospheric levels of ^{14}C, anthropogenic activity has resulted in significant fluctuations in ^{14}C content. The utilization of fossil fuels since the late nineteenth century has introduced ^{14}C- (and ^{13}C-) depleted CO_2 into the atmosphere (the 'Suess effect'). In sharp contrast to this gradual change, nuclear weapons testing in the 1950s and 1960s resulted in the rapid injection of an additional source of ^{14}C into the environment. The amount of ^{14}C in the atmosphere nearly doubled, and the $\Delta^{14}C$ of tropospheric CO_2 increased to greater than $+900$ ppt in the early 1960s. Following the above-ground weapons test ban treaty of 1962, this value has been decreasing as the excess ^{14}C is taken up by oceanic and terrestrial sinks for CO_2. This anthropogenically derived $^{14}CO_2$ 'spike' serves as a useful tracer for the rate at which carbon moves through its global cycle. Any carbon reservoir currently having a $\Delta^{14}C$ value > 0 ppt has taken up some of this 'bomb-^{14}C'. Carbon pools that exhibit no increase in $\Delta^{14}C$ over their 'pre-bomb' values have therefore been isolated from exchange with atmospheric CO_2 during the last 50 years. This contrast between 'pre-bomb' and 'post-bomb' $\Delta^{14}C$ values can serve as an excellent tracer of biogeochemical processes over short timescales. These changes in ^{14}C concentrations can be seen in the updated picture of the Santa Monica Basin regional environment shown in **Figure 2B**, where bomb-^{14}C has invaded everywhere except for the deep basin waters and older sedimentary deposits.

In general, the global distribution of organic ^{14}C is complicated by these interreservoir mixing and exchange processes. The more end-member sources contributing organic carbon to a sample, the more complicated it is to interpret a measured $\Delta^{14}C$ value, especially when trying to translate that value to chronological time. Source-specific ^{14}C dating is needed, and this requires isotopic measurements at the molecular level.

Compound-specific ^{14}C Analysis: Methods

The ability to perform natural-abundance ^{14}C measurements on individual compounds has only recently been achieved. This capability arose from refinements in the measurement of ^{14}C by AMS that allow increasingly small samples to be measured,

and from methods that resolve the complex mixtures encountered in geochemical samples into their individual components. Here we describe the methods that are currently used for this purpose.

Selection of Compounds for ^{14}C Analysis

The organic matter in marine sediments consists of recognizable biochemical constituents of organisms (carbohydrates, proteins, lipids, and nucleic acids) as well as of more complex polymeric materials and nonextractable components (humic substances, kerogen). Among the recognizable biochemicals, the lipids have a diversity of structures, are comparatively easy to analyze by gas chromatographic and mass spectrometric techniques, and are resistant to degradation over time. These characteristics have resulted in a long history of organic geochemical studies aimed at identifying and understanding the origins of 'source-specific' lipid 'biomarker' compounds. Frequently, lipids from several organic compound classes are studied within the same sample (**Figures 3** and **4**).

Although many of the most diagnostic compounds are polar lipids that are susceptible to modification during sediment diagenesis (e.g., removal of functional groups, saturation of double bonds), several retain their marker properties through the preservation of the carbon skeleton (**Figure 5**). Thus sterols (e.g., cholesterol) are transformed to sterenes and ultimately steranes. The isotopic integrity of the compound is also preserved in this way.

It is sometimes the case that families of compounds can also be characteristic of a particular source. For example, plant waxes comprise homologous series of *n*-alkanes, *n*-alkanols, and *n*-alkanoic acids (**Figure 3**). As a result, ^{14}C measurements of a compound class can yield information with similar specificity to single compound ^{14}C analysis, with the benefits of greater total

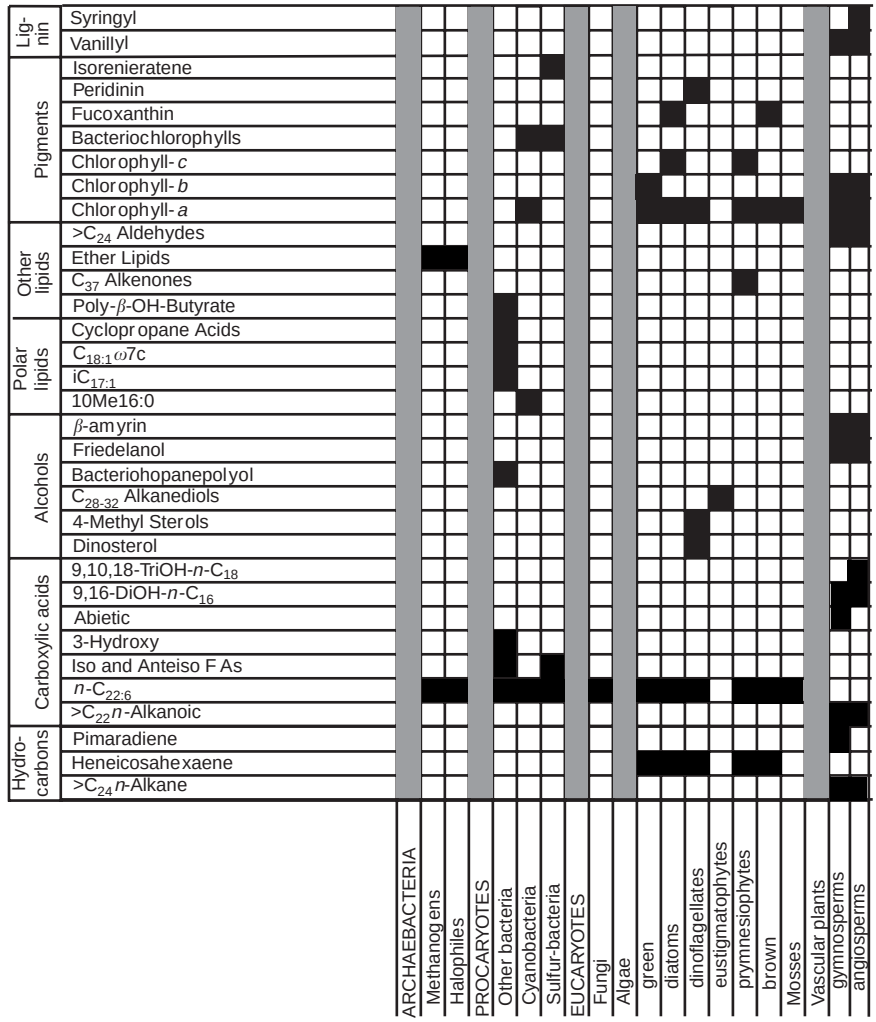

Figure 3 Common source assignments of lipid biomarkers (Modified from Hedges and Oades (1997).)

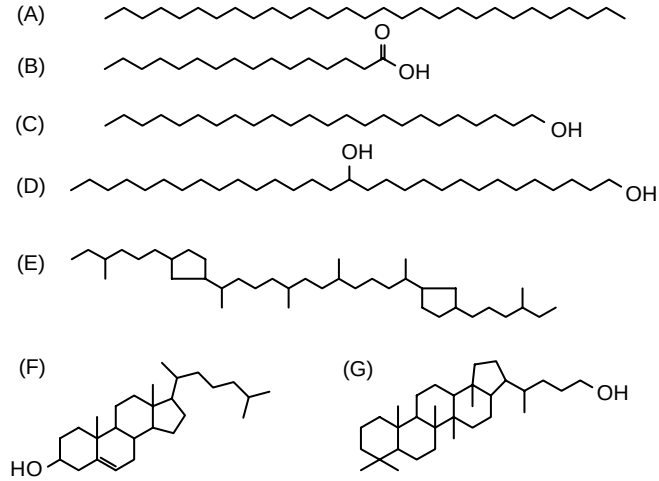

Figure 4 Selected example structures (carbon skeletons and functional groups) for biomarkers shown in Table 1: (A) *n*-C_{29} alkane; (B) *n*-$C_{16:0}$ alkanoic acid; (C) *n*-C_{24} alkanol; (D) C_{30} alkanediol; (E) $C_{40:2cy}$ isoprenoid; (F) C_{27} Δ^5-sterol (cholesterol); (G) C_{32} hopanol.

analyte abundance and, potentially, simpler isolation schemes.

Compound Separation and Isolation

Procedures for single-compound ^{14}C analysis are quite involved, requiring extraction, purification, modification and isolation of the target analytes (**Figure 5**). For lipid analyses, the samples are processed by extracting whole sediment with solvents such as methylene chloride, chloroform, or methanol to obtain a total lipid extract (TLE). The TLE is then separated into compound classes using solid–liquid chromatography. The compound classes elute on the basis of polarity differences, from least polar (hydrocarbons) to most polar (free fatty acids) under normal-phase chromatographic conditions. Individual compounds for ^{14}C analysis are then isolated from these polarity fractions. Additional chromatographic steps or chemical manipulations may be included to reduce the number of components in each fraction prior to single compound isolation, or to render the compounds amenable to isolation by the method chosen. These steps may include silver nitrate-impregnated silica gel chromatography (separation of saturated from unsaturated compounds), 'molecular sieving' (e.g., urea adduction, for separation of branched/cyclic compounds from straight-chain compounds), and derivatization (for protection of functional groups, such as carboxyl or hydroxyl groups, prior to gas chromatographic separation).

For ^{14}C analysis by AMS, tens to hundreds of micrograms of each individual compound must be isolated from the sample of interest. Isolation of individual biomarkers from geochemical samples such as marine sediments and water column particulate matter requires separation techniques with high resolving power. To date, this has been most effectively achieved through the use of automated preparative capillary gas chromatography (PCGC; **Figure 6**).

A PCGC system consists of a commercial capillary gas chromatograph that is modified for work on a semipreparative, rather than analytical, scale. Modifications include a large-volume injection system; high-capacity, low-bleed 'megabore' (e.g., 60 m length $\times$ 0.53 mm inner diameter $\times$ 0.5 μm stationary phase film thickness) capillary columns; an effluent splitter; and a preparative trapping device in which isolated compounds are collected in a series of cooled U-tube traps. Approximately 1% of the effluent passes to a flame ionization detector (FID) and the remaining 99% is diverted to the collection system. The traps are programmed to receive compounds of interest on the basis of chromatographic retention time windows determined from the FID trace. Computerized synchronization of the trapping times permits collection of multiple identical runs (often >100 consecutive injections). Using PCGC, baseline resolution of peaks can be achieved at concentrations >100-fold higher than typical analytical GC conditions, allowing up to 5 μg of carbon per chromatographic peak, per injection, to be separated (to achieve greater resolution, typical loadings are usually about 1 μg of carbon per peak). An example of a typical PCGC separation is shown in **Figure 7**, where ~ 40–130 μg of individual sterols (as their acetate derivatives) were resolved and isolated from a total sterol fraction obtained from Santa Monica Basin surface sediment.

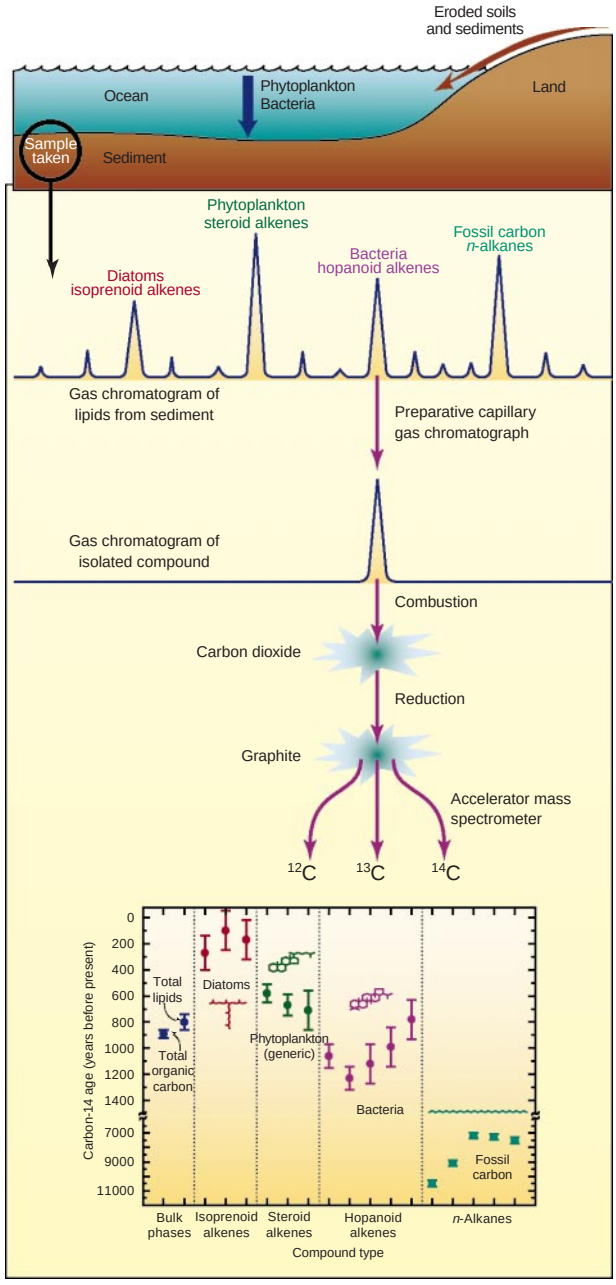

Figure 5 Schematic diagram showing steps for the isolation and ^{14}C analysis of individual sedimentary lipids.

Another practical means of isolating individual components from compound mixtures is high-performance liquid chromatography (HPLC). While the resolving power of HPLC is lower, this technique is particularly suited to polar, nonvolatile, or thermally unstable analytes that are difficult to separate by GC. It also offers higher loading capacity than capillary GC.

In addition to chromatographic resolution and capacity, two additional aspects that require consideration are the potential for contamination of the analytes during the isolation procedure, and corrections for carbon associated with any derivative groups that have been appended to the molecule of interest. Regarding the former, entrainment 'bleed' of chromatographic stationary phase can result in significant carbon contamination of the isolated compound, unless steps are taken to avoid this problem (e.g., use of ultra-low bleed GC columns, removal of contaminants after the chromatographic isolation). This problem is likely to be most acute in HPLC when reversed-phase chromatographic phases are used. Comparison of yields and the $\delta^{13}C$ compositions of the isolated compound and the CO_2 resulting from its combustion are effective means of assessing potential contamination problems.

AMS Measurement of ^{14}C

The purified compounds are sealed in evacuated quartz tubes with CuO as an oxidant. The material is combusted to CO_2, purified, and then reduced to graphite over cobalt or iron catalyst. The mixture of graphite and catalyst is loaded into a cesium sputter ion source. ^{14}C-AMS analysis is performed using special methods necessary for the accurate determination of $\Delta^{14}C$ in samples containing only micrograms, rather than milligrams, of carbon. AMS targets containing $< 150\,\mu g$ of carbon are prone to machine-induced isotopic fractionation, which appears to be directly related to the lower levels of carbon ion current generated by these samples. Therefore, small samples are analyzed with identically prepared, size-matched small standards to compensate for these effects. The f_m values that are calculated relative to these standards no longer show a size-dependent fractionation.

Examples of Applications

Lipid Biomarkers in Santa Monica Basin Sediments

As one example of the application of single-compound radiocarbon analysis, we show a detailed data set for a range of lipid biomarkers extracted from marine sediments. This work focused on the upper few centimeters of a core from Santa Monica Basin. The basin has a high sedimentation rate, and its suboxic bottom waters inhibit bioturbation. As a result, laminated cores recovered from the basin depocenter allow decadal resolution of recent changes in the ^{14}C record. On the timescale of radiocarbon decay, these samples are contemporary and have no in situ ^{14}C decay. However, the $\Delta^{14}C$ values of the end-member carbon sources have

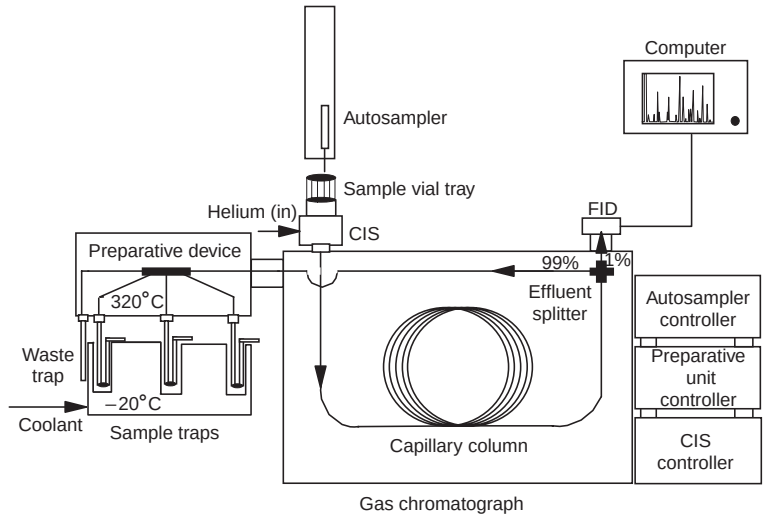

Figure 6 Diagrammatic representation of a preparative capillary gas chromatograph (PCGC) system.

changed (**Figure 2A, B**). 'Bomb-^{14}C' has invaded the modern surface ocean phytoplankton and the terrestrial biota, and through subsequent sedimentation of their organic detritus, this bomb-^{14}C is carried to the underlying sediments. The contrast between 'pre-bomb' and 'post-bomb' Δ^{14}C values, or the relative rate of bomb-^{14}C uptake, therefore is a useful tracer property. It can help distinguish biogeochemical processes that transfer carbon within years or decades (source-specific lipids that now contain bomb-^{14}C) from biogeochemical processes that do not exchange with atmospheric CO_2 on a short timescale (lipids that remain free from bomb-^{14}C).

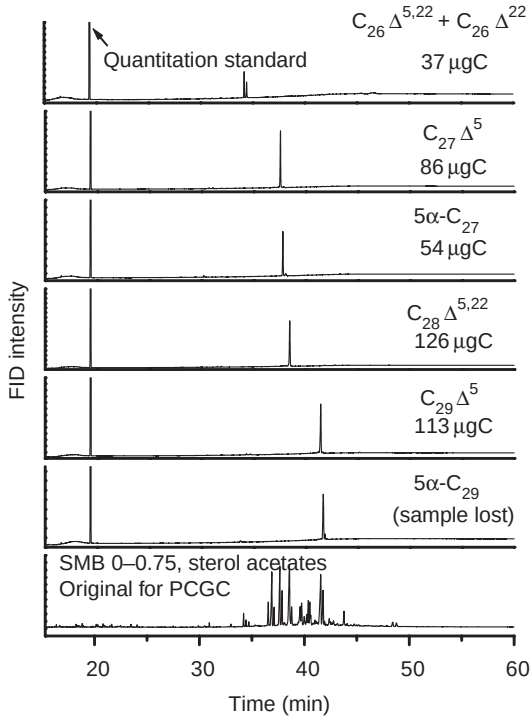

Figure 7 An example PCGC series, showing the total original mixture and the six individual, trapped compounds. In this case the analytes are sterols (as their acetate derivatives). (From Pearson (2000).)

Compound-specific Δ^{14}C values for 31 different lipid biomarker molecules are shown in **Figure 8** for sedimentary horizons corresponding to pre-bomb (before AD 1950) and post-bomb (1950–1996) eras. These organic compounds represent phytoplanktonic, zooplanktonic, bacterial, archaeal, terrestrial higher plant, and fossil carbon sources. The lipid classes include long-chain *n*-alkanes, alkanoic (fatty) acids, *n*-alcohols, C_{30} mid-chain ketols and diols, sterols, hopanols, and C_{40} isoprenoid side chains of the ether-linked glycerols of the *Archaea*.

The data show that the carbon source for the majority of the analyzed biomarkers is marine euphotic zone production. Most of the lipids from 'pre-bomb' sediments have Δ^{14}C values equal to the Δ^{14}C of surface water DIC at this time (dotted line), while most of the lipids from 'post-bomb' sediments have Δ^{14}C values equal to the Δ^{14}C of present-day surface water DIC (solid line).

However, it is clear that two of the lipid classes do not reflect carbon originally fixed by marine photoautotrophs. These are the *n*-alkanes, for which the Δ^{14}C data are consistent with mixed fossil and contemporary terrestrial higher plant sources, and the archaeal isoprenoids, for which the Δ^{14}C data are consistent with chemoautotrophic growth below the euphotic zone. This is just one example of the way in which compound-specific ^{14}C analysis can distinguish carbon sources and biogeochemical processes simultaneously. The large number of compounds that appear to record the Δ^{14}C of surface

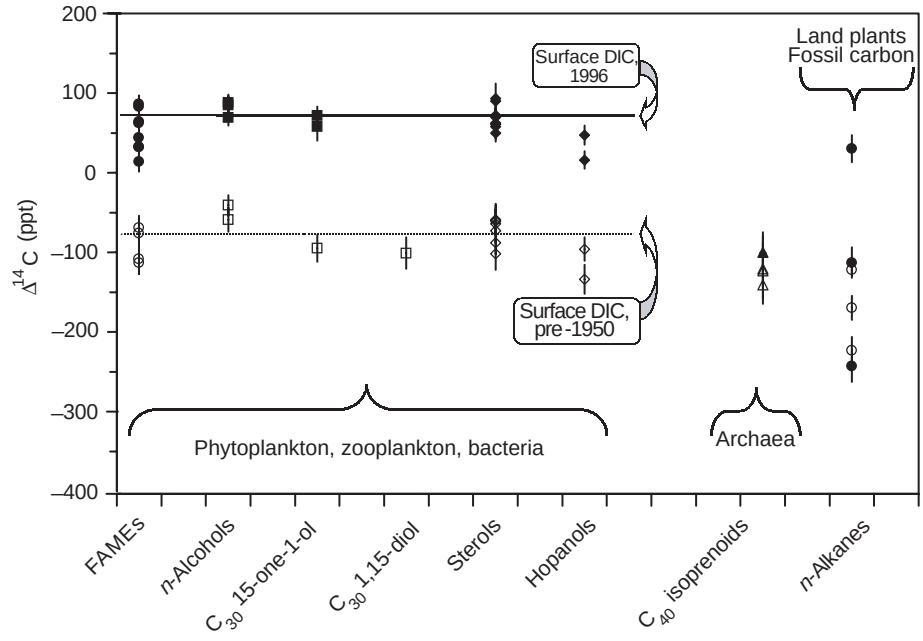

Figure 8 Δ^{14}C data for individual lipids extracted from Santa Monica Basin sediments. The solid symbols represent compounds extracted from the post-bomb sedimentary horizon (AD 1950–1996). The hollow symbols represent compounds extracted from the pre-bomb sedimentary horizon (deposited prior to AD 1950). (Modified after Pearson (2000).)

water DIC, and therefore marine primary production, points to the potential for numerous tracers of marine biomass; these are the target compounds of interest when developing refined sediment chronologies. In particular, the sterols appear to be particularly effective tracers of surface ocean DIC, and hence suitable for this purpose.

Monosaccharides in Oceanic High-Molecular-Weight Dissolved Organic Matter

The second example illustrates the utility of single compound, as well as compound class, ^{14}C measurements as ocean process tracers. In this case, the process of interest is the cycling of dissolved organic matter (DOM) in the ocean. Much progress has been made in characterizing this large carbon pool. A significant fraction of the DOM pool is composed of high-molecular-weight (HMW) compounds (>1 kDa), and a substantial fraction of this HMW DOM is known to be comprised of complex polysaccharides. Evidence suggests that these polysaccharides are produced in the surface ocean as a result of primary productivity, and/or attendant heterotrophic activity, and should therefore carry a bomb-influenced ^{14}C signature. Similar polysaccharides have been detected in HMW DOM well below the surface mixed layer, implying that these compounds are transported to the deep ocean. Two possible mechanisms can explain these observations: (1) advection of DOM associated with ocean circulation, and/or (2) aggregation and vertical transport followed by disaggregation/dissolution at depth. Because the timescales of aggregation and sinking processes are short relative to deep water formation and advective transport, ^{14}C measurements on polysaccharides in HMW DOM provide means of determining which mechanism is dominant.

Figure 9 shows vertical ^{14}C profiles for DIC and DOM as well as ^{14}C results for selected samples of sinking and suspended particulate organic matter (POM), HMW DOM, and monosaccharides isolated from selected depths at a station in the North-east Pacific Ocean. Individual monosaccharides were obtained by hydrolysis of HMW DOM, and purified and isolated by HPLC. The similarity of Δ^{14}C values of individual monosaccharides implies that they derive from a common polysaccharide source. Such similarities in ^{14}C lend support to the utility of ^{14}C measurements at the compound class level. Furthermore, the similarity between Δ^{14}C values of these compounds and surface ocean DIC indicates that they are either directly or indirectly the products of marine photoautotrophy. The deep ocean (1600 m) data shows the presence of bomb-radiocarbon in the monosaccharides. Their enrichment in ^{14}C relative to DIC, and similarity to suspended POC at the same depth, suggests that this component of HMW DOM is

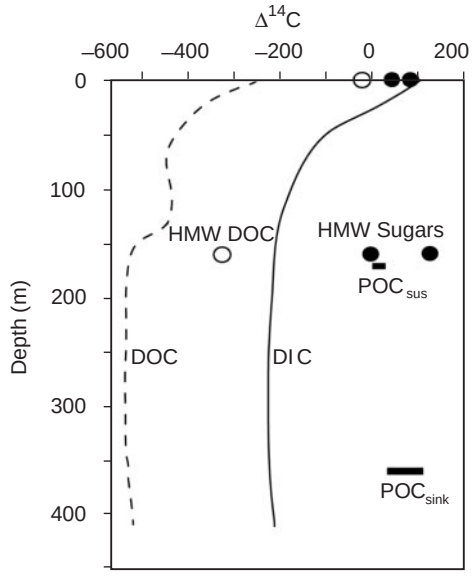

Figure 9 $\Delta^{14}C$ values (ppt) for different fractions of carbon in the North-East Pacific: solid bars labeled POC_{sus} and POC_{sink} correspond to suspended and sinking POC, respectively; solid and dashed lines show depth profiles for DIC and DOC, respectively; open circles are total HMW DOC and closed circles are individual sugars (monosaccharides) isolated from the HMW DOC fractions. (Modified from Aluwihare (1999).)

injected into the deep ocean by vertical transport as particles.

Summary

The ability to perform single-compound ^{14}C measurements has only recently been realized, and as a consequence its application as a tracer in ocean sciences remains in its infancy. The above examples highlight potential applications of single-compound ^{14}C measurements as tools for understanding the biogeochemical cycling of organic matter in the ocean. There are several other areas of study where this approach holds great promise. For example, ^{14}C measurements of vascular plant biomarkers (e.g., plant waxes, lignin-derived phenols) in continental shelf sediments provide constraints on the timescales over which terrestrial organic matter is delivered to the ocean. The 'infinite ^{14}C age' signature that polycyclic aromatic hydrocarbons and other fossil fuel-derived contaminants carry provides an effective means of tracing their inputs to the coastal ocean relative to contributions from natural processes (e.g., biomass burning). As methods are streamlined, it is anticipated that single compound ^{14}C measurements will find increasing application in marine biogeochemistry.

See also

Ocean Carbon System, Modelling of. Radiocarbon. Stable Carbon Isotope Variations in the Ocean.

Further Reading

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