BOTTOM LANDERS

R. A. Jahnke, Skidaway Institute of Oceanography, Georgia, USA

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Introduction

The need to understand better the chemical, biological and transport processes across the deep sea floor has fueled major engineering advances in sea floor instrumentation in the last few decades. In principle, the kinds of experiments that can be conducted on the sea floor are limited only by researchers' imaginations. In reality, the majority of the instruments that have been developed seek to accomplish two types of basic operations: deploy, sample and retrieve benthic flux chamber systems to directly measure sea floor exchange rates and deploy *in situ* sensor arrays capable of measuring the vertical distribution of solutes in near-surface pore waters from which exchange rates can be estimated based on pore-water transport models. Generically, the instrument frames used to deploy these types of instruments have been called 'Bottom Landers'. This term was coined in the early 1970s in recognition of the similarity between the approximate shape and function of these devices and the more famous 'Lunar Lander' that carried the first men to the moon. In the following, the design strategies and basic instrumentation for conducting benthic flux chamber incubations and sensor measurements at the deep sea floor are discussed.

Benthic Flux Measurement Strategies

There are two major strategies for estimating benthic fluxes. Benthic flux chamber incubations can be performed in which a known volume of bottom water is trapped above a known area of seafloor. Any solute transported out of the sediments under the chamber will be trapped within the chamber waters. Hence, the concentration of this constituent within the chamber waters will increase with incubation time. Conversely, the chamber water concentration of any chemical constituent that is being transported into the sediments will decrease with incubation time. The seafloor flux is directly proportional to the rate of concentration increase or decrease, the volume of bottom water and area of the seafloor enclosed by the chamber. The major strength of this method is that there is minimal disturbance to the sediment system, maximizing the probability of accurate estimates. In addition, the fluxes obtained will reflect the net exchange due to all transport processes occurring within the spatial and temporal scales of the chamber incubations. A weakness of this approach is that, other than inferences about total integrated reaction rates within the sediment column supporting the observed benthic flux, little information is gained about the reaction processes and distributions themselves.

The other major strategy for estimating seafloor fluxes is to measure the concentration gradient of chemical constituents very near (preferably across) the sediment–water interface. Knowing the transport processes, the flux can be calculated based on the transport rates and measured concentration gradients. In principle, this method can be applied to all types of transport processes. In practice, however, the exact nature of nondiffusive transport processes is unknown and this calculation strategy is used almost exclusively to estimate exchange due to molecular diffusion only. Thus, a limitation of this approach is that accurate benthic fluxes can not be estimated in a location where pore water exchange due to processes other than molecular diffusion is significant. On the other hand, a strength of using concentration profiles is that the distribution of reactions can be assessed and inferences of reaction mechanisms can be made from the variations in the fluxes estimated at different depths below the sediment surface.

In the 1960s and 1970s, it was increasingly recognized that the temperature and pressure changes that occur in bringing a sample from the deep seafloor to the sea surface may cause a variety of changes in the sample itself. Examples include changes in metabolic rates of benthic populations, changes in pore water concentration gradients from which diffusive fluxes are calculated and changes in chemical concentrations in pore waters due to pressure or temperature driven reactions. Additional artifacts continue to be identified to this day. Thus, analyzing samples brought to the deck of a ship has been increasingly recognized as being not accurate enough to study important questions such as: what are the respiration rates of sea floor populations? and what is the seafloor dissolution rate of calcium carbonate? Conducting benthic flux chamber incubations and pore water concentration profile measurements directly on the seafloor, at *in situ* temperature and pressure, is one strategy for avoiding sampling artifacts and improving the accuracy of deep ocean measurements.

Benthic Chamber Landers

Design and Operations

The development of benthic chamber landers followed the experiences gained from conducting chamber or 'bell jar' incubations in shallow water areas where they could be tended by SCUBA divers. The major challenge in developing this instrumentation was simply to automate the required operations to function reliably in the harsh conditions of the deep seafloor. The earliest instrument that was routinely deployed in the deep sea and has provided a large data set is the free vehicle grab respirometer (FVGR) developed in the late 1970s by Dr K. L. Smith, Jr at Scripps Institution of Oceanography in San Diego, California, USA (**Figure 1**).

The basic instrument consists of a structural frame upon which the critical components are mounted. To minimize weight, the frame is most often constructed of tubular aluminum. Instrument flotation is mounted on the upper portion of the instrument frame. The most common type of flotation is glass spheres (as shown in Figure 1) although syntactic foam flotation has also been used. The latter is much more expensive than glass spheres but is not susceptible to implosion, which is critical if submersibles are ever required to work in close proximity. Both types provide relatively constant buoyancy. At the top of the flotation section are mounted devices such as a flag, strobe light, radio transmitter, or satellite transmitter to help locate the instrument when it is at the sea surface.

Expendable weights are mounted at the lower portion of the tubular frame, usually adjacent to the 'feet'. As these instruments are free vehicles, these weights provide the negative buoyancy necessary to drive the instrument to the seafloor. A latching mechanism permits the weights to be released by acoustic command from a surface vessel at the end of the experiment. Upon release, the flotation provides sufficient buoyancy to raise the instrument back to the sea surface where it may be recovered by a surface research vessel.

The specific instrument packages, controlling electronics and other operational components are mounted in the central part of the frame. It is in these components where the greatest differences

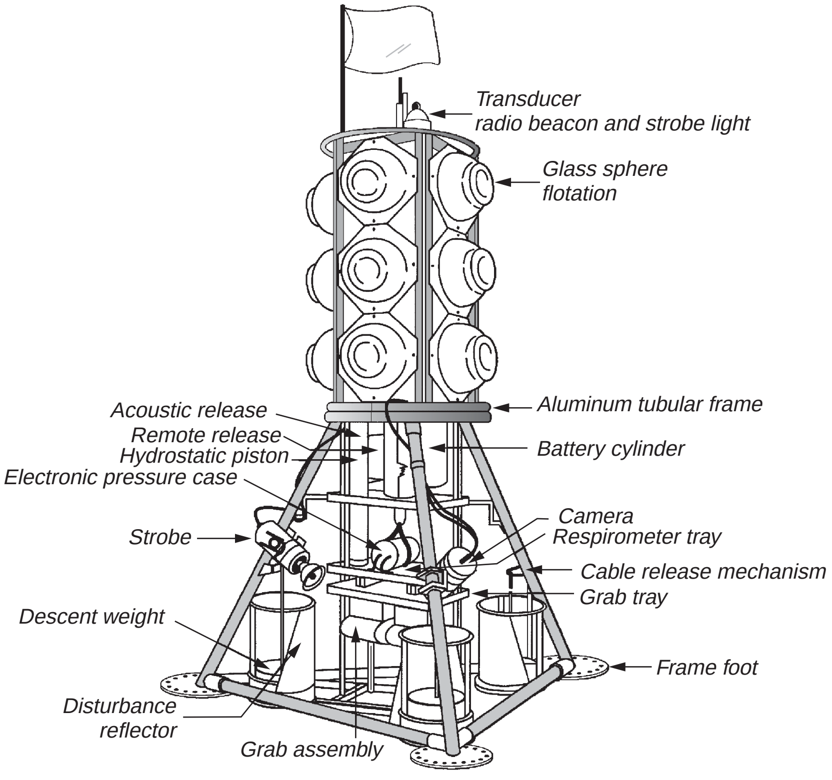

Figure 1 Schematic of the Free Vehicle Grab Respirometer developed by Dr Kenneth L. Smith, Jr. (Scripps Institution of Oceanography). Instrument height is approximately 4 m.

between individual designs occur. Chamber instruments have been designed to accommodate one to four chambers simultaneously. Although increasing desirable replication, increased numbers of chambers on a single instrument tend to increase variability in the data by decreasing chamber area and increasing the size and complexity of the control, data storage and sampling systems.

Early chamber systems like the FVGR relied on oxygen polarographic electrodes to quantify the changes in oxygen within the chambers. Thus, the early instruments were only capable of estimating benthic oxygen demand. Although electrodes continue to be widely used, many modern instruments also employ an electronically controlled sampling system so that benthic exchange of many types of solutes can be assessed.

The utility of these instruments is demonstrated by their proliferation. Today, there are more than 20 different research groups that have developed *in situ* benthic flux chamber instruments. These groups have designed and implemented numerous modifications and alterations to Smith's initial design, yet the basic characteristics and capabilities of the chamber lander remain the same. Examples of the numerous important modifications that have been made to the early designs include several types of stirring mechanisms and time-series sampling systems so that the exchange of solutes other than oxygen can be addressed. In addition to the basic benthic flux chamber operations, complementary sampling devices have also been added. For example, one chamber instrument design has incorporated an *in situ* whole core squeezer. This device recovers a sediment core and then sequentially squeezes the surface pore waters from the sediments to provide high resolution pore water samples. Although not appropriate for all solutes due to surface exchange during squeezing, these samples provide important information concerning the pore water gradients of selected metabolites, such as oxygen and nitrate, that greatly enhances the interpretation of the flux chamber results.

Examples of results from a benthic flux chamber deployment at approximately 3000 m on the continental rise of the eastern US seaboard are shown in **Figure 2**. On each plot, the concentration is on the vertical axis and the horizontal axis represents incubation time. These results demonstrate the range of possible responses. Solutes that are taken up by the sediments tend to decrease with incubation time. An example of this is oxygen which is consumed in the sediments through benthic respiration. Most other components are produced in the sediments through the decomposition or dissolution of biogenic debris. This production supports a flux out of the sediments and the concentrations of these constituents increase with incubation time. Specific examples of these include phosphate, that is released from degrading organic tissue, total inorganic carbon (TIC) that is released through respiration and the dissolution of $CaCO_3$, titration alkalinity (TA) that is produced by the dissolution of $CaCO_3$ and silicate which is produced by the dissolution of opal. Some species, such as nitrate, may increase or decrease with incubation time depending on the relative rates of the competing sedimentary processes that produce or consume them. For example, nitrate is produced by the oxidation of ammonium (nitrification) and is consumed by denitrifiying bacteria below the oxic zone. Whether there is a flux out of or into the sediments depends on the ratio of these rates.

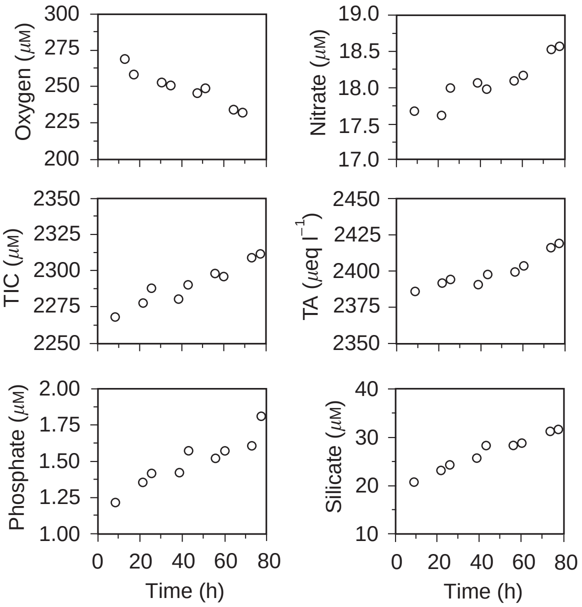

Figure 2 Example benthic flux chamber results from 2927 m on the continental rise of the US eastern seaboard using the chamber system of Dr Richard A. Jahnke (Skidaway Institute of Oceanography). Measurement precision for each analysis is: oxygen ± 1.5%, nitrate, phosphate and silicate ± 2%, TA ± 0.5%, TIC ± 1%.

Direct estimates of benthic fluxes can be made from these results from the relationship below.

$$\text{Benthic flux} = (S \times V)/A$$

where S is the slope of the concentration vs. time results; V is the volume of the bottom water trapped in the chamber; A is the sediment surface enclosed by the chamber.

Note that if the chamber has vertical sides, V/A = height of the water column trapped within the chamber.

Major Design Controversies and Differences

Despite the numerous advantages of benthic flux incubations for estimating seafloor exchange, there are a variety of limitations and concerns that need to be addressed in evaluating results from chamber incubations and in assessing the relative merits of the different instrument designs.

As shown in the example results, concentration changes are required to evaluate the solute exchange rate. However, the concentration changes will also alter the near-surface gradients, altering the flux. In the extreme case, where concentration changes are large, such as complete oxygen depletion within the chamber, changes in chamber water chemistry may alter near-surface chemical reactions and/or exchange processes, greatly altering the chemical flux. To minimize this source of uncertainty, chamber deployments must be designed to minimize the concentration changes and maintain the chamber sediments as near to their natural state as possible. Thus, it is important that the most precise analytical procedures be employed so that accurate fluxes can be quantified without large concentration changes.

Maintaining natural benthic fluxes requires that the sediment surface not be disturbed during chamber deployment. This need has resulted in several different types of deployment strategies and instrument designs. Most instruments have been designed to settle to the seafloor and then, after some preset time period, slowly insert the chambers into the bottom. It is hoped that this waiting period will allow the sediments that are resuspended by the impact of the instrument with the bottom to be swept away, leaving a natural surface on which to conduct the incubation. Of course, if the ‘bow wave’ or disturbance ahead of the descending instrument is sufficiently large, the entire sediment surface within the perimeter of the instrument frame will be impacted. Recognizing this, some instrument designs positioned the chambers lower than the main frame so that the chamber would encapsulate the sediments upon which the flux incubation will be performed before the bow wave from the frame reaches the surface. Another approach is to suspend a descent weight 10 m below the instrument. This weight provides the negative buoyancy needed for the instrument to descend to the seafloor. Once the weight reaches the sea floor, the positive buoyancy of the instrument itself causes it to remain suspended above the bottom. A hydraulic winch system is then activated and the instrument package is very slowly pulled to the seafloor. A recent innovative approach for minimizing bottom disturbance has been implemented on the ROVER lander. This benthic flux chamber device is capable of ‘crawling’ around the seafloor. Thus, once the instrument has impacted the bottom, it simply crawls laterally to an undisturbed location prior to conducting benthic flux chamber incubations and pore water profiling operations.

Perhaps the most controversial aspect of benthic flux incubations and instrument design is focused on mechanisms and requirements for simulating natural flow conditions within the chamber. Exchange across the sediment surface requires transport across the hydrodynamic boundary layer including the molecular diffusive sublayer. Hydrodynamic conditions within the chamber control the thickness of this layer. Changes in the hydrodynamic regime will alter the thickness of the layer and, at least temporarily, will alter the exchange rate. However, since the benthic flux is supported by metabolic and chemical reactions in the sediments that are not influenced by the diffusive sublayer thickness, the seafloor exchange rate will eventually revert to its natural rate. Thus, the need to accurately reproduce the natural hydrodynamic conditions with the benthic flux chamber depends critically on the response time of the surface pore water gradients and the length of the chamber incubations. If deployments are short relative to the gradient response times, accurate fluxes will require the maintenance of near-natural hydrodynamic conditions within the chamber. On the other hand, if the deployments are long relative to the pore water gradient response times, the fluxes will be insensitive to chamber hydrodynamics and a simple chamber water stirring mechanism is adequate. For most solutes in the deep sea, deployments greater than 10–20 hours are sufficient to minimize artifacts due to changes in the diffusive sublayer thickness and thus only require a relatively simple mixing strategy.

Sensor Landers

Design and Operations

The basic instrument design and field deployment operations of the sensor landers are the same as that already discussed for the benthic flux chamber. The instrument consists of a frame, identical to that of a benthic flux chamber lander, with flotation mounted at the top and expendable descent weights attached near the feet. The major difference is that the benthic chamber is replaced with an instrument package capable of inserting microelectrodes with high vertical resolution into the surface sediments. An example of the basic instrument package is presented in **Figure 3**.

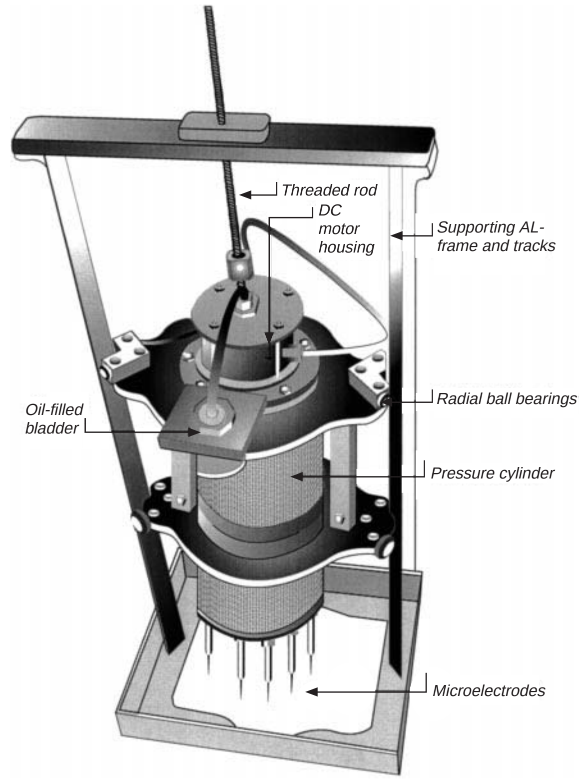

Figure 3 Schematic of the microelectrode deployment apparatus developed by Dr Clare Reimers (Rutgers University). Instrument pressure case is approximately 50 cm.

Each of these types of packages consists of three basic components: the sensing electrodes themselves; a mechanism for moving the electrodes vertically across the sediment–water interface; and the data storage and controlling electronics. In the design pictured in **Figure 3**, the sensing electrodes attached directly to the main pressure case. Located within the pressure case are the power and electronics necessary to operate the motor and electrodes and to provide for data storage and retrieval. The motor drive system is positioned outside the case to move the case vertically so that profiles of the measured components are obtained. The vertical resolution of the profiles is controlled by the size of the electrode tip and the precision of the motor-drive assembly.

Early designs primarily employed oxygen and electrical resistivity electrodes. The latter is required to assess the porosity and tortuosity of the sediments. These characteristics influence the rate of diffusion in sediments. In recent years, numerous other sensors have been developed and implemented for deep sea lander use. These include pH, PCO_2, total CO_2, calcium, ammonium, and nitrate. It is anticipated that the development of other sensors will continue and the types of measurements possible in the future will continue to grow.

An example of oxygen, resistivity (formation factor) and pH results obtained from a sensor lander deployment is provided in **Figure 4**. Because oxygen is consumed within the sediments, primarily due to the respiration by benthic organisms but possibly also due to chemical consumption, oxygen concentrations decrease with increasing depth in the sediments. This downward concentration gradient implies a benthic flux into the sediment. Contrastingly, pH first increases and then decreases with sediment depth. This more complicated profile shape is due to competing reactions downcore. In this example, pH first increases with depth due to

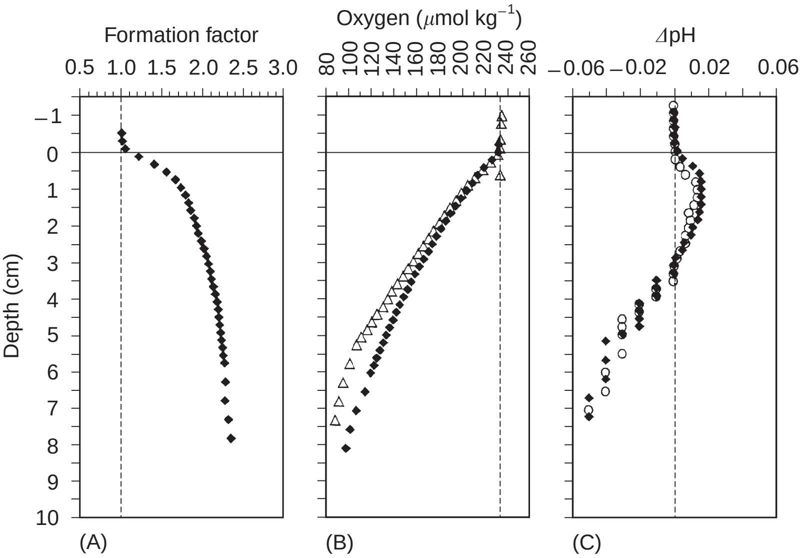

Figure 4 Examples of deep seafloor (A) resistivity, (B) microelectrode oxygen and (C) pH results obtained by Dr Burke Hales (Oregon State University).

the dissolution of calcium carbonate and then decreases with depth due to continued production of carbon dioxide.

Formation factor values increase with sediment depth due to the compaction of sediment particles. This decrease in resistivity implies a decrease in sediment void space (porosity) and increase in the path length required for solutes to travel around the particles (tortuosity). These factors tend to decrease the effective diffusion coefficient with depth in the sediments.

These results can be used to estimate the diffusive flux across the sediment–water interface and to evaluate the reaction rate within the measured depth interval. The benthic flux can be estimated from Fick's First Law corrected for porosity and the effects of sediment particles on diffusion rates as shown below.

$$\text{Benthic flux} = \phi D_s \delta C/\delta z$$

where ϕ is surface porosity; D_s is effective sediment diffusion coefficient; $\delta C/\delta z$ is vertical concentration gradient at the sediment–water interface.

Porosity near the sediment surface is generally estimated from a regression of the measured resistivity and directly measured porosities at wider-spaced depth intervals. The latter are generally determined from the weight loss upon drying bulk sediments. The effective diffusion coefficient can be estimated by dividing the molecular diffusion coefficient by the porosity and tortuosity (i.e. $D_s = D/(\phi\theta)$, where D is molecular diffusion in free solution, and θ is tortuosity). This expression requires independent measurements of porosity and tortuosity at the same vertical scale as the concentration profile. These measurements are not always available. A common alternative, but less accurate, strategy in fine-grained sediments is to approximate the effective diffusion coefficient as the molecular diffusion coefficient times the square of the porosity. This relationship has been empirically derived from numerous individual studies.

It is important to note that the flux equation shown above is not limited to the sediment surface but rather can be used to evaluate the diffusive flux at any depth horizon within the profile. Thus, it is often useful to define a sediment layer of a small thickness and calculate the diffusive fluxes at the top and bottom of the layer. Assuming that horizontal diffusive exchange can be neglected and that the profile is in steady state, the difference in these fluxes is a measure of the net production or consumption rate of the measured solute within the layer. Thus, by interpreting the vertical variations in the flux, one can evaluate the distributions of reactions in the sediments and potentially make inferences about the processes and mechanisms controlling solute diagenesis and benthic flux.

Advantages, Limitations and Design Concerns

Unlike benthic flux chambers that require a significant incubation time, sensor landers can obtain a profile relatively quickly, usually within 1–2 hours. The exact required time is determined by the number of sampling depths and the response time of the sensors employed. Thus, sensor landers can be used relatively rapidly to obtain *in situ* profiles and estimate diffusive fluxes at the deep seafloor.

There are also several limitations to this approach. Because the measurements are generally made within several hours of the instrument reaching the seafloor, the accuracy of the flux estimate can be severely compromised by physical disturbance of the sediment surface. Potential disturbance can be caused by the instrument itself or by the bow wave that precedes the instrument as it settles. Video recordings of these instruments settling onto the bottom reveal significant resuspension of surface sediments. Although the resuspended sediments are generally allowed to settle back to the seafloor or be advected away prior to making measurements, the effect of this disturbance on the profiles is still a concern. The only instrument to completely avoid this potential problem is the ROVER lander.

Maintaining natural hydrodynamic conditions within the benthic boundary layer is also critical to the accuracy of microelectrode flux estimates. Unlike benthic flux chamber incubations that extend over time intervals sufficient to return to initial conditions if diffusive boundary layer thicknesses are altered, electrode measurements are rapid and would record transient conditions if made directly after altering bottom hydrodynamic conditions. Because the instrument frame and electrodes themselves may alter bottom flow, such changes are a concern.

Since the electrode sensing tips are very small (generally 5–20 μm in diameter) as required to achieve fine vertical resolution, they also respond to horizontal variations that may be caused by burrowing organisms or physical inhomogeneities. Because the geochemical questions being asked often require knowledge of the mean benthic flux for a known area or region, numerous profiles are often required to estimate the average profile and benthic flux.

Special Landers

In addition to the types of landers discussed above, numerous landers have been developed in the last

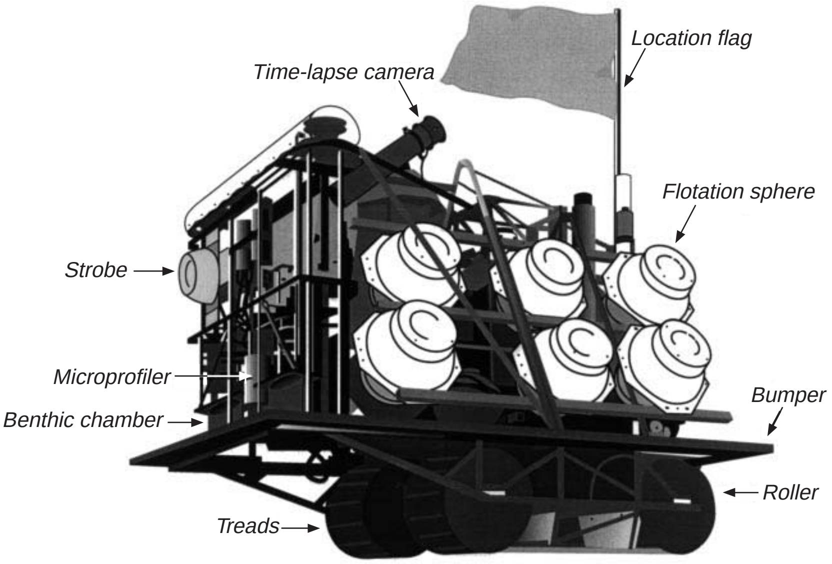

Figure 5 The ROVER Lander developed by Dr Kenneth L. Smith, Jr (Scripps Institution of Oceanography). Instrument is approximately 2 m in height and 3 m in length.

decade for special purposes and it is likely that new types will continue to be developed in the future. For example, simple benthic flux chamber landers have been developed to measure advective pore water flows around hydrothermal vent and mid-ocean ridge systems. In these types of chambers, osmotic pumps are used to continuously add a tracer and remove a sample. Pore water advection rates as low as 0.1 mm year^{-1} can be measured with this system. For seafloor microbial studies, a lander has been developed capable of injecting radiolabeled tracers continuously throughout the upper 70 cm of the sediment column. The sediments surrounding the line of injection are cored and recovered at the end of a preset incubation period and returned to the ship for analysis.

Perhaps the most innovative recently developed lander is the ROVER (**Figure 5**). This instrument is capable of performing repeated benthic flux chamber incubations and microelectrode profiling measurements at depths as great as 6000 m. Most importantly, after a measurement cycle is complete, the instrument uses a tractor tread propulsion system to move approximately 5 m so that the next measurement is performed on a natural, undisturbed surface. Since this instrument is crawling laterally on the sediment surface, it eliminates the potential disturbances discussed for the free fall instruments. This instrument was constructed to examine the temporal variations in seafloor fluxes and is capable of performing duplicate benthic flux chamber and microelectrode profiling measurements at 30 individual sites over a 6-month period on a single deployment.

Conclusions

In conclusion, bottom lander technology has been developed in the last several decades so that it is now possible to perform a wide variety of sampling procedures and incubation experiments remotely on the deep seafloor. These capabilities have greatly improved our understanding of the benthic processes at abyssal depths.

See also

Benthic Boundary Layer Effects. Benthic Organisms Overview. Deep-sea Fauna. Deep-sea Sediment Drifts. Pore Water Chemistry.

Further Reading

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BOTTOM WATER FORMATION

A. L. Gordon, Columbia University, Palisades, NY, USA

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Introduction

Meridional sections of temperature and salinity through the Pacific and Atlantic Oceans (**Figure 1**) reveal that in the Pacific below 2000 m, more than half of the ocean depth, the water is colder than 2°C. The Atlantic is somewhat warmer, but there too the lower 1000 m of the ocean is well below 2°C. Only within the surface layer, generally less than the upper 500 m of the ocean is the water warmer than 10°C, amounting to only 10% of the total ocean volume. The coldness of the deep ocean is due to interaction of the ocean with the polar atmosphere. There, surface water reaches the freezing point of sea water. Streams of very cold water can be traced spreading primarily from the Antarctic along the sea floor, warming en route by mixing with overlying water, into the world's oceans (**Figure 2**).

The coldest bottom water, Antarctic Bottom Water (AABW), is derived from the shores of Antarctica. There, freezing point, high oxygen concentration, water is produced during the winter over the continental shelf. At a few sites the shelf water salinity is sufficiently high, greater than 34.61‰, that, on cooling to the freezing point, the surface water density is sufficiently high to allow it to sink to great depths of the ocean. As the shelf water descends over the continental slope into the deep ocean it mixes with adjacent deep water, but this water is also quite cold so the final product arriving at the seafloor at the foot of Antarctica is about − 1.0°C. Definitions used by different authors vary, but generally AABW is defined as having a potential temperature (the temperature corrected for adiabatic heating due to hydrostatic pressure) less than 0°C. AABW spreads into the lower 1000 m of the world ocean, where it cools and renews oxygen concentrations drawn down by oxidation of organic material within the deep ocean. AABW is said to ventilate the deep ocean.

In the Atlantic Ocean the 2°C isotherm marks the base of a wedge of relatively salty water, associated with high dissolved oxygen and low silicate concentrations (see **Figure 1**). This water mass is called North Atlantic Deep Water (NADW). The densest component of NADW is formed as cold surface waters during the winter in the Greenland and Norwegian Seas. This water sinks to fill the basin north of a ridge spanning the distance from Greenland to Scotland. Excess cold water overflows the ridge crest, mixing on descent with warmer more saline water, producing a bottom water product of about + 1.0°C. The overflow water stays in contact with the sea floor to near 40°N in the Atlantic Ocean, where on spreading southward it is lifted over the remnants of denser AABW.

Export of Greenland and Norwegian Sea bottom water has been estimated from a series of current measurements. Transports of about $2 \times 10^6\,\mathrm{m^3\,s^{-1}}$ of near 0.4°C water occur between the Faroe Bank and Scotland, $1 \times 10^6\,\mathrm{m^3\,s^{-1}}$ of similar water passes through notches between Iceland and Faroe Bank, and $3 \times 10^6\,\mathrm{m^3\,s^{-1}}$ of near 0°C is exported through the Denmark Strait, between Greenland and Iceland. The overflow plumes rapidly entrain warmer waters, producing bottom water of near + 1.0°C. With entrainment of other deep water, a production rate of about $8 \times 10^6\,\mathrm{m^3\,s^{-1}}$ of overflow water is likely. Less dense components of NADW, that do not contact the seafloor are formed in the Labrador Sea and Mediterranean Sea. The total production of NADW is estimated as $15 \times 10^6\,\mathrm{m^3\,s^{-1}}$.

As the Antarctic is the primary source of the cold bottom waters of the world ocean, Antarctic Bottom Water is discussed in this article. *See* **North Atlantic Deep Water** for further information on that Northern Hemisphere deep water mass.