Accelerator mass spectrometry in geologic research

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ABSTRACT

The ability of accelerator mass spectrometry (AMS) to measure very small concentrations of the nuclides ^{10}Be, ^{14}C, ^{26}Al, ^{36}Cl, and ^{129}I has led to many innovative applications in geologic research. To take advantage of this opportunity in the geosciences, it is important to understand how AMS works, how these nuclides are produced, and how they can be applied to geologic problems. We first discuss the basics of AMS, explaining what gives the method its ability to count small numbers of these nuclides. We review how these nuclides are produced and transported in the atmosphere, hydrosphere, and lithosphere. We then explain the ways that AMS is being used to solve a wide range of problems in geologic research by discussing specific applications in areas such as geomorphology, tectonics, climatology, hydrology, and geochronology.

Keywords: accelerator mass spectrometry, cosmogenic nuclides, geochronology, surface-exposure dating, isotope geology.

INTRODUCTION

Accelerator mass spectrometry (AMS) is becoming increasingly important in geologic research, as well as in archaeology, biomedicine, and environmental science (Tuniz et al., 1998; Fifield, 1999). This paper provides an overview of AMS techniques, their strengths and limitations, and a brief review of some geologic applications.

The strength of AMS lies in its ability to count small numbers of certain nuclides whose production and subsequent behavior can be reconstructed to yield valuable information concerning geologic processes. Two other common techniques to measure nuclides are decay counting and conventional mass spectrometry. In general, these two more established techniques are less expensive and easier to use than AMS; however, for certain nuclides, AMS is the only method with the sensitivity required to measure the small concentrations found in natural samples (Elmore and Phillips, 1987; Finkel and Suter, 1993; Vogel et al., 1995).

The main nuclides of geologic interest currently measured by using AMS, which we refer to as ''AMS nuclides,'' are listed in Table 1. Their half-lives range from 5730 yr to 16 m.y. Nuclides with much shorter half-lives are more efficiently measured by simpler nuclear-decay counting techniques. In fact, ^{14}C (half-life of 5730 yr) is a borderline case; for this carbon isotope, both AMS and decay counting are in use. On the other hand, nuclides with very long half-lives (i.e., $>$100 m.y.) are not so rare because they have not completely decayed since the origin of the solar system. When the nuclide is not rare, normal mass spectrometry suffices for counting (Vogel et al., 1995).

In most cases, geologic work involving AMS uses cosmogenic nuclides. These are nuclides produced in the atmosphere (''meteoric'') or in rocks (''in situ'') by cosmic rays. Anthropogenic nuclides, those produced by human activities, are also sometimes used as tracers. For example, the atmospheric nuclear weapons tests in the 1950s and early 1960s produced large amounts of AMS nuclides such as ^{14}C and ^{36}Cl (Synal et al., 1990; Levin and Kromer, 1997). Finally, in some cases, production in rocks as a by-product of natural radioactivity can be important (Sharma and Middleton, 1989).

Many significant advances in geologic research have been made possible by the use of AMS. These include (1) establishing glacial chronologies by dating moraines, erratics, and glacially polished surfaces (Zreda and Phillips, 1995); (2) studying past behavior of the climate and the Earth's magnetic field by measuring the ^{10}Be content of ice and sediment cores (Frank, 2000); (3) radiocarbon dating a single, small component of a soil sample (such as the pollen grains) in order to obtain a well-defined date (Brown et al., 1989); (4) using ^{10}Be to establish the presence of subducted sediment in magma produced at island arcs (Monaghan et al., 1988); (5) determining basin-scale erosion rates by measuring in situ–produced cosmogenic nuclides (Granger et al., 1996); (6) dating tectonic events by the accumulation of cosmogenic nuclides in freshly exposed fault scarps or offset surfaces (Mitchell et al., 2001); (7) measuring river incision rates by dating fluvial terraces (Hancock et al., 1999); and (8) determining erosion rates of exposed bedrock surfaces (Bierman and Caffee, 2002).

This paper provides an introduction for geologists with no previous knowledge of AMS. For readers who want a more detailed discussion of AMS and its applications in the earth sciences, good general references include the book by Tuniz et al. (1998) and the review papers by Elmore and Phillips (1987), Finkel and Suter (1993), Vogel et al. (1995), Fifield (1999, 2000), and Gosse and Phillips (2001).

AMS METHODOLOGY

In many ways, AMS is similar to conventional mass spectrometry. Both techniques measure isotopic ratios in samples that are first ionized, then accelerated by an electric field, and then separated by mass and charge in a magnetic field. The key difference in AMS is that the ions are accelerated to a high energy (usually on the order of tens of millions of electron volts) by using a tandem ac-

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TABLE 1. MAIN AMS NUCLIDES USED IN GEOLOGY

Nuclide	Half-life	Target material (negative ion)	Ratio measured (interfering isobars)	Important atmospheric production mechanisms	Important lithospheric production mechanisms near surface
^{10}Be	1.5 m.y.	BeO (BeO$^-$)	^{10}Be/^{9}Be (^{10}B)	Spallation on N and O	Spallation on Si and O (in quartz)
^{14}C	5730 yr	Graphite (C$^-$)	^{14}C/^{13}C or ^{14}C/^{12}C	Neutron reaction with N	Spallation on Si and O (in quartz)
^{26}Al	720 k.y.	Al_2O_3(Al$^-$)	^{26}Al/^{27}Al	Spallation on Ar	Spallation on Si (in quartz)
^{36}Cl	301 k.y.	AgCl (Cl$^-$)	^{36}Cl/^{35}Cl or ^{36}Cl/^{37}Cl or ^{36}Cl/[^{35}Cl + ^{37}Cl] (^{36}S)	Spallation on Ar	Spallation on K and Ca; neutron capture on Cl
^{129}I	15.7 m.y.	AgI (I$^-$)	^{129}I/^{127}I	Spallation on Xe	Spontaneous fission of ^{238}U

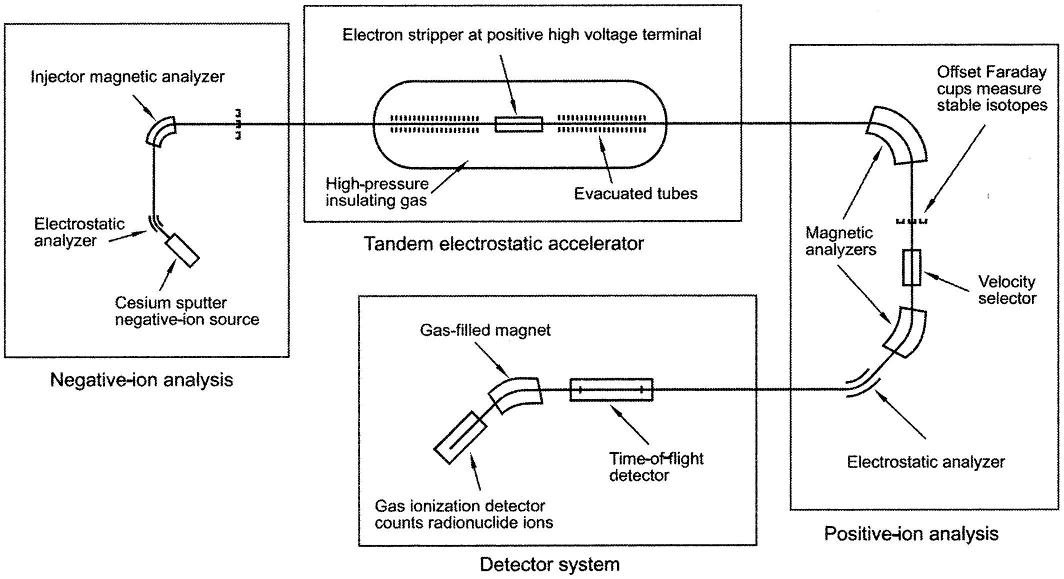

Figure 1. Diagram of an AMS system (adapted from Elmore and Phillips, 1987). The unstable, rare nuclides travel from the ion source to the ionization detector, while the stable, abundant nuclides are counted in the Faraday cup after the accelerator. The figure includes components that are used in only some AMS systems, such as the gas-filled magnet, the time-of-flight detector, the second electrostatic analyzer, and the third magnetic analyzer. In some systems a solid-state detector (Tuniz et al., 1998) is used instead of the ionization detector. The nuclides travel down an evacuated tube, and the accelerator tank is filled with an insulating gas to prevent electrical discharges.

celerator; at this high energy, several important techniques for discriminating against unwanted nuclides become effective. These techniques allow one to reject the unwanted nuclides without reducing the counting rate for the nuclide of interest to unacceptably low levels.

Figure 1 shows the typical components of an AMS system. After chemical preparation, the sample is placed in the ion source where the element of interest is negatively ionized. The goal is to use the various electric and magnetic fields to guide the nuclide of interest along the beam line while rejecting unwanted nuclides and molecules. The nuclide of interest reaches the detector and is counted while all other nuclides either never make it to the detector or are recognized at the detector as an unwanted signal.

It is very difficult to count the absolute number of atoms of a particular nuclide in the sample. Only a fraction of the atoms of the element of interest are turned into negative ions in the ion source. Also, ions are inevitably lost in the long journey between the ion source and the detector. So, in AMS, one always measures the ratio of two numbers, the number of ions of the nuclide of interest, and the number of ions of a corresponding stable isotope. For example, ^{26}Al ions and ^{27}Al ions are counted alternately by changing the electric and/or magnetic fields to obtain the ratio ^{26}Al/^{27}Al. Table 1 lists the stable partner measured with each unstable nuclide.

This measured ratio, ^{26}Al/^{27}Al, does not precisely reflect the true ratio in the sample because the two isotopes may not be ionized or transmitted through the system in exactly the same way. To solve this problem, one also measures the ratio for a known, standard material and then makes an appropriate renormalization of the ratio for the unknown sample. At present AMS can measure ratios as small as 10^{-16} to 10^{-14} for the isotope pairs used in geologic research.

In some cases, the ratio of the two isotopes is actually the desired result. For example, for radiocarbon dating the ratio ^{14}C/^{13}C is what one needs to compute a date (McNichol et al., 2001). In most cases, however, the ratio is not the actual quantity of interest. For example, when one is using ^{26}Al, the concentration of ^{26}Al in the sample is the quantity of interest. The ^{27}Al concentration must be measured separately and is combined with AMS measurement of the ^{26}Al/^{27}Al ratio to compute the concentration of ^{26}Al. In some cases (Al and C), the stable isotope occurs naturally in the sample, whereas in other cases (i.e., Be), it may be added to the sample as a spike.

Preparation of the Sample

Before an AMS measurement can be made, the sample must be prepared. The nuclides of interest must be extracted from their geologic matrix and converted into a form suitable for AMS analysis. (General aspects of sample preparation are covered in Finkel and Suter [1993], Tuniz et al. [1998], and Gosse and Phillips [2001].) This step often involves considerable time and expense and plays a crucial role in the success of the measurement.

The goals of sample preparation are (1) to prepare a chemical form that will optimize the formation of negative ions in the ion source, (2) to remove unwanted isobars (nuclei with the same total number of protons and neutrons, such as ^{14}C and ^{14}N; their mass difference is very small) from the sample, (3) to minimize contamination of the sample, and (4) to minimize preparation time and cost given requirements 1–3.

Passage of the Sample Through the AMS System

The prepared sample is then placed in the ion source; the purpose of the ion source is to produce negative ions either of the nuclide under study or of a molecule containing the nuclide. For example, C^- ions are produced for radiocarbon studies. However, because Be has a very weak tendency to form negative ions, the molecule BeO^- is used.

A cesium sputter source is most commonly used for AMS (Middleton, 1983; Finkel and Suter, 1993; Tuniz et al., 1998; Southon and Roberts, 2000). The sample is in solid form, and Cs^+ ions are accelerated to 3–15 keV before striking the sample. Atoms are sputtered from the sample, some in the form of negative ions. Cesium also serves to coat the surface of the sample, which lowers the work function and thus makes it more likely to produce a negative ion. An ion source is designed to maximize two related, but distinct quantities: (1) the intensity of the appropriate negative-ion beam, and (2) the efficiency, which is the fraction of the desired atoms in the sample that is ionized and fed into the beam.

At present, the currents produced by a well-designed ion source, for the abundant stable isotope, range from 1 to 100 μA measured before the accelerator (Southon and Roberts, 2000). To get an idea of what these numbers mean, consider a case in which the current of $^{9}BeO^-$ is 10 μA and in which a sample has a $^{10}Be/^{9}Be$ ratio of 10^{-14}. The current of $^{10}BeO^-$ ions is then 10^{-19} A, which corresponds to ~36 ions per minute.

After leaving the ion source, the negative ions are accelerated to 40–100 keV by an electric field and then passed through the magnetic analyzer (Fig. 1). The ions are then accelerated to a high energy (on the order of several million electron volts) by the first half of the tandem Van de Graaff electrostatic accelerator. In such an accelerator, the central region (the terminal) is kept at a high positive potential with respect to either end. Thus, the negative ions are accelerated to the terminal. The ions then pass through a gas or a foil stripper that removes several electrons from the ion, making a positive ion that is then further accelerated by the second half of the electrostatic accelerator, as the positive ion is repelled by the terminal. This electron stripping also serves to break up any molecules present in the beam.

The positive ions are now at a high energy, on the order of tens of millions of electron volts. They pass through the magnetic analyzers (Fig. 1) and usually through an electrostatic analyzer and/or Wien filter as well. The stable, abundant isotope (e.g., ^{13}C or ^{27}Al) is not counted in the nuclear detector because the current of this isotope is so high it would overwhelm the detector. Instead, the stable isotope is fed into a Faraday cup, which measures the electrical current. The rare isotope is counted one atom at a time, usually by a gas-ionization detector.

The AMS system discriminates against unwanted nuclides in several ways (Finkel and Suter, 1993; Tuniz et al., 1998; Fifield, 1999):

1. The electric fields are arranged at the ion source so that only negative ions are transported to the injector magnet. Thus, atoms that do not form negative ions are rejected at this very first stage. For example, ^{14}N is the common isotope of nitrogen and has a mass very close to that of ^{14}C. However, ^{14}N does not form a negative ion, and so it can reach the injector magnet only as part of a molecular ion. Other examples of unwanted isobars (and the ions of interest with which these unwanted isobars could interfere) that do not form negative ions are ^{26}Mg (^{26}Al), ^{36}Ar (^{36}Cl), and ^{129}Xe (^{129}I).

2. The injector magnet rejects unwanted nuclides in the second stage of the AMS system. Only ions having a certain value of ME/q^2 (where M is the mass, E is the kinetic energy, and q is the charge) are correctly deflected and continue to the next stage. At this point, $q = -1$, and the kinetic energy E is set by the previously applied accelerating potential.

3. After the negative ions have been accelerated in the first half of the tandem accelerator, they pass through the stripper, which removes several electrons, causing molecules to dissociate. The ionic molecule $^{12}CH^-_2$, for example, has a mass very close to that of ^{14}C and so is passed through the injector magnet along with ^{14}C ions. It is broken up by the stripper, and the fragments are removed by the next discrimination stages.

The stripper produces a distribution of positive ions in different charge states; the number of ions in each charge state is dependent on the voltage and the characteristics of the stripper gas or foil. For example, to attain maximum efficiency, energies on the order of 3 MeV are needed for C^{3+}, and energies on the order of 7 MeV are needed for Cl^{7+}. Subsequent discrimination stages select only one of these charge states, typically comprising 10%–50% of the ions.

4. After leaving the accelerator, the ions pass though the analyzing magnet, which again selects a particular value of ME/q^2. The values of E and q have been changed since the ion passed through the injector magnet; if the ion entered the accelerator as part of a molecule that was subsequently broken up in the stripper, its value of M is also significantly different.

5. Fragments of molecules broken up by the stripper, and ions that change charge at places other than the terminal, may not have the assumed value of the kinetic energy E and so may accidentally have the correct value of ME/q^2 and not be rejected by the analyzing magnet. An electrostatic analyzer (which selects for E/q), and/or a Wien (velocity) filter (which selects for E/M), must be used to remove these particles, whereas a time-of-flight detector can be used to identify them. In addition, another magnetic selection is often inserted, either just before or just after this stage.

6. The last stage in discrimination occurs when the rare nuclides are counted, usually in a silicon solid-state detector or a gas-ionization detector. Because a gas detector allows measurement of the rate of energy loss of the ion and because the rate of coulombic energy loss depends on the value of Z (the nuclear charge), one can reject isobars of the desired nuclide as long as their intensity is not too great. For example, when counting ^{36}Cl, the ^{36}S survives all previous stages and arrives at the detector. However, the ^{36}S ($Z = 16$) loses its kinetic energy at a slower rate than ^{36}Cl ($Z = 17$), allowing a computer to distinguish their energy-loss spectra. For this type of detector to be effective, the ions must have a high enough energy, on the order of tens of millions of electron volts. The detector counts one ion at a time and is ''dead'' for a short time while it is registering either a sulfur or a chlorine ion, so too much sulfur in the sample causes the detector to be dead most of the time and unavailable to count the ^{36}Cl ions. A gas-

filled magnet (in which the ions travel in a gas-filled chamber as they go through the magnetic field region) can be used to prevent unwanted isobars from reaching the detector, as explained in Fifield (1999). When measuring ^{10}Be, a metal foil of appropriate thickness can be placed in front of the detector; this foil will stop most of the ^{10}B, whereas most of the ^{10}Be will reach the detector.

COSMIC RAYS AND THE PRODUCTION OF NUCLIDES

In most cases, the AMS nuclides studied in geologic research are produced in the cascades of particles initiated by galactic cosmic rays. Less energetic solar cosmic rays make a minor contribution (Lal and Peters, 1967; Lal, 1991; Masarik and Reedy, 1995; Masarik and Beer, 1999; Lal, 2000; Gosse and Phillips, 2001). The galactic cosmic rays are made up mostly of protons, with some heavier nuclei mixed in. They have energies up to ~10 GeV. As these primary cosmic rays travel through the atmosphere, they lose energy through ionization of molecules and through nuclear interactions with atomic nuclei. These nuclear interactions produce new particles called secondaries, which if they are energetic enough, will travel through the atmosphere and produce yet more secondaries. The secondary-particle cascade includes photons, pions, electrons, muons, neutrons, and neutrinos (Lal and Peters, 1967).

The cosmogenic nuclides studied by AMS are produced mainly by neutrons, protons, and muons (Tuniz et al., 1998). If their energy is high enough, protons and neutrons can collide with a nucleus and cause a spallation reaction. In a spallation reaction, the target nucleus is greatly disrupted; the result of the reaction is a somewhat smaller nucleus and several other emitted particles. Another production mechanism involves the capture by a target nucleus of a slow neutron or negative muon (Phillips et al., 2001; Heisinger et al., 2002b). Finally, fast muons can initiate cascades of particles that then interact with target nuclei (Heisinger et al., 2002a).

The interaction of muons with nuclei is much weaker than the interactions of protons and neutrons (Lal and Peters, 1967). Because of this fact, muons penetrate much farther into the lithosphere than protons or neutrons (Heisinger et al., 1997). Thus, production by proton and neutron spallation is the dominant mechanism in the atmosphere (Tuniz et al., 1998). At the Earth's surface, energetic protons are very rare, so in situ production is due mainly to neutrons (Kurz and Brook, 1994). At greater depths, total production is reduced and is due mostly to muons (Heisinger et al., 1997).

Cosmogenic Nuclide Production

In the atmosphere and the lithosphere, the production rate of cosmogenic nuclides at a particular location depends on the following factors (Lal and Peters, 1967; Finkel and Suter, 1993; Masarik and Reedy, 1995; Tuniz et al., 1998; Masarik and Beer, 1999; Lal, 2000; Gosse and Phillips, 2001):

Geomagnetic Latitude

The Earth's magnetic field deflects particles with low energy away from the Earth. This deflection is strongest at low geomagnetic latitudes (where the magnetic field lines are mostly perpendicular to the velocity of incoming particles) and becomes less significant at high geomagnetic latitudes, resulting in a latitude dependence of the production rate of AMS nuclides (Rose et al., 1956). For example, production due to spallation at sea level at the equator is ~60% of its value at sea level at 60° latitude.

Altitude

As cosmic-ray particles pass through and react with the molecules of the atmosphere, the number and energy spectrum of the particles change. In addition, the density of target nuclei is a function of altitude. This results in a strong dependence of production rate on altitude (for a useful discussion, see Desilets and Zreda, 2001). For example, at 40° latitude, the production rate in rocks due to spallation is about two times greater at an altitude of 800 m than at sea level.

Solar Activity

The inner solar system is permeated by a magnetic field generated by the charged particles emitted by the Sun. This magnetic field deflects some of the galactic cosmic rays away from the Earth, resulting in a production rate that varies inversely with solar activity (Lal, 1988; Masarik and Beer, 1999; Beer et al., 2002).

Time

The effects discussed so far all depend on time; thus, the production rate at a given location is not a constant, but varies on a wide range of time scales. For example, Earth's magnetic field varies in magnitude and direction on a variety of time scales (Wagner, 1998; Guyodo and Valet, 1999; Frank, 2000). The location of the geomagnetic poles wanders on scales of hundreds of years (Wagner, 1998); there is strong evidence that the overall magnitude of the field has changed significantly over the past 50 k.y. (Guyodo and Valet, 1999; Frank, 2000). Finally, at widely spaced intervals the magnetic field completely reverses its polarity; the last such event happened at ca. 780 ka (Cande and Kent, 1995). The level of solar activity also varies with time. Solar activity oscillates over the 11 yr sunspot cycle, and there is evidence for variations over longer time scales as well (Beer et al., 1994, 2002).

In Situ Production

For most dating and erosion research involving in situ nuclides, the production rate is integrated over the accumulation period for the nuclide, with however, a bias toward more recent time. This bias arises in two ways: (1) nuclear decay removes a smaller portion of the more recently produced nuclides; (2) on an eroding landform, a given clast moves towards the surface as time goes on, and so the more recent production rate is less attenuated by depth. This integration effectively averages over short-term production variations and lessens the effects of time variations in the production rate (Clark et al., 1995; Dunai, 2001; Masarik et al., 2001).

For some research, one needs to know the production rates as a function of depth beneath the surface (Lal, 1991; Heisinger et al., 1997; Stone et al., 1996; Gosse and Phillips, 2001). Production due to neutron spallation decays exponentially with depth, decreasing by a factor of e^{-1} every 0.6 m in rocks of density 2.6 g/cm^3. Production from thermal neutrons, as in the neutron-capture reaction that yields ^{36}Cl from ^{35}Cl, is sometimes important; this mechanism has a complicated depth dependence, peaking at a depth of ~0.3 m and then decaying exponentially with a length scale of ~0.5 m (Phillips et al., 2001). Finally, the production due to muons, although a small effect at the surface, becomes dominant at depths of a few meters (Stone et al., 1998; Heisinger et al., 2002a, 2002b).

For any particular research involving in situ produced nuclides, the investigator needs to know the production rates at the location of interest. Although the factors that determine this rate are understood, it is quite difficult to perform a theoretical calculation that would accurately predict the production for a given location and a given target mineral or rock. Thus, much experimental research has been devoted to this issue: (1) Production rates at specific locations have been measured for various nuclides; for example, one can measure the ^{26}Al and ^{10}Be produced in quartz by cosmic rays at a location for which the exposure time has been determined by some other method (see, e.g., Nishiizumi et al., 1989; Kubik et al., 1998). (2) A laboratory-generated beam of particles can be used to generate AMS nuclides in a particular target (see, e.g., Heisinger et al., 1997, 2002a, 2002b). (3) Scaling formulas have been developed, so that if the production rate at a particular latitude and altitude is known, the rate at some other latitude and altitude can be determined. These scaling equations have been largely empirical in nature, and there is still active research devoted to understanding this scaling and to achieving agreement among the various proposed scaling equations. (Lal, 1991; Desilets and Zreda, 2001; Dunai, 2001; Gosse and Phillips, 2001; Stone, 2000).

SPECIFIC COSMOGENIC NUCLIDES

^{10}Be

Atmospherically produced ^{10}Be is studied in ice cores and sediment (terrestrial, lacustrine, and marine). In situ produced ^{10}Be is usually studied in quartz; work has been done on ^{10}Be in olivine (Seidl et al., 1997) and in pyroxene (Ivy-Ochs et al., 1998). There is still uncertainty in the value of the ^{10}Be half-life; for a discussion, see Gosse and Phillips (2001).

Cosmic rays interact with nitrogen and oxygen in the atmosphere to make ^{10}Be. An estimate of the global atmospheric production rate is ~10^6 atoms per square centimeter per year (Finkel and Suter, 1993). Beryllium has a residence time of ~1 yr in the stratosphere and a few weeks in the troposphere. It then attaches to aerosol particles and is deposited on Earth's surface (Baumgartner et al., 1997b).

In quartz, ^{10}Be is produced directly by spallation reactions with silicon and oxygen nuclei. At depths greater than ~3 m below the surface, production by muons becomes dominant (Heisinger et al., 2002a, 2002b). At sea level and high latitude, the production rate at the surface is ~5 atoms per year per gram of quartz (for a discussion of the conflicting estimates for this rate, see Gosse and Phillips, 2001). An advantage of using quartz is that it is a very tight matrix; the ^{10}Be atoms produced in the quartz are not likely to escape, and the ^{10}Be atoms produced in the atmosphere will stay on the surface of the quartz grains (Lal and Arnold, 1985). One of the problems with measuring ^{10}Be produced in whole-rock samples is that such contamination is often difficult to eliminate (Gosse and Phillips, 2001).

In ice cores or sediment, there can be several sources for the ^{10}Be, and care must be exercised in sample processing to extract the desired component (Gu et al., 1996; Baumgartner et al., 1997b). For in situ quartz studies, pure quartz is separated from the raw sample material by using physical and chemical means (Kohl and Nishiizumi, 1992; Tuniz et al., 1998). Methods include magnetic separation, density separation, and etching with HF. The quartz is then dissolved in HF.

Because beryllium is a rare element, a sample usually contains very little of the stable isotope ^{9}Be; thus it is necessary to add 0.1–1.0 mg of ^{9}Be to make enough sample material for AMS. After adding the carrier, the beryllium is isolated by ion chromatography and selective precipitation and then combusted to form BeO.

^{14}C

The ^{14}C nuclide is produced in the atmosphere by cosmic rays through the interaction of neutrons with nitrogen. It is rapidly oxidized to form CO_2 and attains a uniform distribution throughout the atmosphere. The production rate is ~6×10^7 atoms per square centimeter per year (Finkel and Suter, 1993).

In typical applications, carbon is extracted from organic matter for radiocarbon dating. Carbon can also be extracted from seawater, groundwater, and minerals. In most AMS laboratories, the sample is physically and chemically cleaned, combusted to form CO_2, and then reduced to graphite (Tuniz et al., 1998; McNichol et al., 2001). Great care must be taken to avoid contaminating the sample with extraneous carbon.

The principal advantage of using AMS rather than decay counting for ^{14}C is that with AMS, it is possible to use a much smaller sample. This is because AMS directly counts the ^{14}C atoms, whereas with decay methods the ^{14}C atoms are counted only when they decay. A typical AMS sample contains ~1 mg of carbon, and samples containing only tens of micrograms of carbon are possible; with decay counting, a typical sample requires several grams of carbon.

The graphite is used in the sputter source to produce C^- ions. It is possible to construct an ion source that directly uses the CO_2 gas (Bronk and Hedges, 1990; Bronk Ramsey and Hedges, 1997). Sample preparation is simpler, but gas sources produce a lower current than graphite sources.

Besides its use in traditional radiocarbon dating, a new application for ^{14}C is being developed (Handwerger et al., 1999; Lal and Jull, 2001). The ^{14}C nuclide is produced in situ in rocks by cosmic rays. Efforts are under way to devise a method to extract ^{14}C from the rock and form a suitable target for AMS measurement. One line of research focuses on ^{14}C production in quartz (Lifton et al., 2001); the production rate, at sea level and high latitudes, is estimated to be ~15 ^{14}C atoms per gram of quartz per year.

^{26}Al

At present this nuclide is studied in quartz, where it is produced in situ by cosmic rays. Nuclear spallation reactions on silicon are the main mechanism, although deep enough below the surface, muon reactions become more important. At sea level and high latitude, the surface production rate is ~30 atoms per year per gram of quartz (Gosse and Phillips, 2001).

Sample preparation steps are very similar to those used in processing ^{10}Be in quartz (Kohl and Nishiizumi, 1992; Tuniz et al., 1998). The quartz is purified by chemical and physical means and then dissolved. The aluminum is isolated by ion chromatography and selective precipitation and then combusted to form Al_2O_3.

There are several important points concerning ^{26}Al analysis. (1) The isobar ^{26}Mg does not form a negative ion, so this isobar does not interfere with the measurement of Al^-. (2) Aluminum in the form of the stable isotope ^{27}Al is reasonably abundant in many geologic contexts, and so for most materials the ^{26}Al/^{27}Al ratio is too small to measure; quartz typically contains little aluminum, and this is one reason that ^{26}Al is mainly studied in quartz. (3) The Al^- ion is not very readily formed, so a major challenge for AMS measurements is to produce enough beam for good statistical precision. (4) The AlO^- molecular ion is more easily formed, but the MgO^- ion would interfere.

^{36}Cl

The geologic archives for the ^{36}Cl nuclide are the hydrosphere and glacial ice, where ^{36}Cl is deposited after being produced by cosmic rays in the atmosphere, and rocks, where ^{36}Cl is directly produced by cosmic rays. Atmospheric production is due to cosmic rays interacting with argon. The global production rate is ~3×10^4 atoms per square centimeter per year (Finkel and Suter, 1993). Because chlorine is hydrophilic, the ^{36}Cl produced in the atmosphere tends to be incorporated into water droplets and reaches the surface in precipitation.

In rocks, ^{36}Cl is produced in several ways (Zreda and Phillips, 1994; Stone et al., 1996, 1998; Gosse and Phillips, 2001). Spallation reactions involving ^{39}K and ^{40}Ca produce ^{36}Cl. In addition, neutron capture by ^{35}Cl is an important mechanism; such neutrons can arise either as cosmic-ray secondaries or as a by-product of the decay of uranium and thorium nuclei in the rock. Muon reactions, such as muon capture by ^{40}Ca, can also produce ^{36}Cl. At present there is significant disagreement as to the values of the production rates for these different channels (Gosse and Phillips, 2001). One set of values for sea level and high latitude suggests that the production rate due to calcium spallation is ~49 atoms per year per gram of Ca, the rate due to muon capture on calcium is ~5 atoms per year per gram of Ca, and the rate due to potassium spallation is ~190 atoms per year per gram of K (Stone et al., 1996).

Computation of the production rate for a given rock sample requires a detailed chemical analysis of the rock. Besides knowing the concentrations of ^{35}Cl, ^{39}K, and ^{40}Ca, one must know the concentrations of elements such as gadolinium (Gd), which have high neutron absorption cross sections. Programs are available that compute the production rate of ^{36}Cl, given the chemical composition and geometry of a rock (see Phillips and Plummer, 1996).

The goal of sample preparation is to extract the chlorine from water or rocks and to form AgCl (Bierman et al., 1995; Stone et al., 1996; Tuniz et al., 1998). If one is studying ^{36}Cl produced in situ in rocks, it is important to remove the atmospherically produced ^{36}Cl that may be attached to the exterior of the clasts and grains.

^{129}I

The ^{129}I nuclide is produced by cosmic-ray interactions with atmospheric xenon and as a fission product of ^{238}U in rocks (Elmore et al., 1980; Finkel and Suter, 1993). These two mechanisms lead to an ^{129}I/^{127}I ratio in the hydrosphere of $\sim10^{-12}$ (Moran et al., 1998). Its main application so far has been in hydrologic and oil migration studies (Liu et al., 1997).

The chemical preparation involves extracting the iodine from the water and eventually forming AgI (Tuniz et al., 1998). The isobar ^{129}Xe does not form a negative ion and so does not interfere in the AMS measurement. The biggest challenge is making a clean separation of the isotope ^{127}I in the AMS measurement (Fifield, 1999).

EARTH SCIENCE APPLICATIONS

Surface-Exposure Dating and Erosion Rates

Surface-exposure dating is the most common application of AMS in the earth sciences (Gosse and Phillips, 2001). In the simplest case (i.e., no erosion or inherited nuclides), a rock that has previously been shielded from cosmic rays is suddenly exposed at the surface. As time passes the rock accumulates an inventory of cosmogenic nuclides through in situ production; the number of such nuclides is a measure of how long the rock has been exposed to cosmic rays. This method thus dates the event that exposed the rock.

The AMS nuclides commonly used for surface-exposure dating are ^{10}Be, ^{26}Al, and ^{36}Cl. The nuclides ^{10}Be and ^{26}Al are almost always studied in quartz. ^{36}Cl is more versatile, although its production rate is more uncertain; it can be used in pure mineral phases such as K-feldspar and in whole-rock samples. The nuclide ^{14}C is also being developed for exposure dating; it is produced in situ in minerals such as quartz and calcite by cosmic rays (Handwerger et al., 1999; Lal and Jull, 2001; Lifton et al., 2001).

In addition to the unstable AMS nuclides, we note that the stable nuclides ^{3}He and ^{21}Ne are also used for surface-exposure dating; these are measured by using conventional mass spectrometry. One advantage of using the unstable AMS nuclides is that because they decay, it is more likely that the freshly exposed rock starts with a very small initial amount of the nuclide in question. An advantage of the stable nuclides ^{3}He and ^{21}Ne is that after exposure, the buildup of stable nuclides continues without hitting a saturation limit. In contrast, for the unstable nuclides, the concentration approaches its saturation value for exposure times greater than about two half-lives, and the dating becomes difficult. However, in the great majority of cases the erosion rate is high enough so that nuclear decay is not the limiting factor in the buildup of cosmogenic nuclides.

To derive an exposure age by using a particular nuclide, one must know the production rate for that nuclide at the location (i.e., altitude and latitude) of interest and for the rock or mineral under study. One of the key factors limiting accuracy in surface-exposure dating is that local production rates are uncertain by at least 5%–10%.

One complicating feature of surface-exposure dating is that as the exposed rock surface accumulates cosmogenic nuclides, it is usually also eroding. The concentration of cosmogenic nuclides then depends on the erosion history as well as on the exposure time (Lal, 1991; Gosse and Phillips, 2001). This dependence on the erosion history is an unwelcome complication if the researcher is only interested in the exposure time; however, many workers turn this dependence to their advantage and use the cosmogenic nuclides to make quantitative determinations of rock erosion rates (Lal, 1991; Cerling and Craig, 1994; Small et al., 1999; Gosse and Phillips, 2001; Bierman and Caffee, 2001; Schaller et al., 2001; Bierman and Caffee, 2002).

In some cases, geomorphic evidence, such as the presence of glacial striations, can be used to rule out any significant erosion (Nishiizumi et al., 1989). In cases in which erosion has occurred, the usual approach is to model the erosion process in some way (Lal, 1991; Small et al., 1999; Bierman et al., 1999); for example, one may assume that the erosion rate has been constant since the time of exposure and then calculate the effect such erosion would have on the nuclide concentrations (Lal, 1991). Unless the erosion rate can be established by other means, there are then at least two unknowns involved (the exposure time and the erosion rate) when analyzing cosmogenic nuclide data (Lal and Arnold, 1985; Lal, 1991).

One proposed solution to this problem is to use two nuclides; use of ^{26}Al and ^{10}Be in quartz has been discussed by many authors (see Gillespie and Bierman, 1995). In principle, if we measure the concentrations of both of these nuclides in quartz at the exposed surface, we may derive an exposure time and an erosion rate. However, this method has not proved useful in practice. If the erosion rate is not low enough, the fact that the half-lives of ^{26}Al and ^{10}Be are different will not matter. Typical uncertainties in both the AMS measurements and in the production rates then lead to very large uncertainties in the inferred erosion rate and exposure time. The method has the best chance of working when the erosion rate is small (so that the effective exposure time of a given clast is not limited to a small value by erosion) and the exposure time is long.

It is useful to note that for nuclides produced mainly by spallation reactions, such as ^{26}Al and ^{10}Be, if one simply computes a surface-exposure age by assuming that the erosion rate is zero, the age thus obtained is a lower bound to the true exposure age, if the rock had no inherited cosmogenic nuclides. For example, if the exposure ages are computed in this way for many clasts on an eroding moraine, the oldest one may approximate the age of the surface (Zreda and Phillips, 1995).

Surface-exposure methods have been used to determine chronologies in a variety of situations: volcanic eruptions (see, e.g., Zreda et al., 1993), meteor impact craters (Phillips et al., 1991), rock art (see, e.g., Phillips et al., 1997), glacial advances and retreats (see, e.g., Gosse et al., 1995), and alluvial fans (Van der Woerd et al., 1998). These methods have also been used to establish erosion rates in different geologic contexts. In the following subsections we discuss a few of these applications in more detail.

Glacial Chronology

Establishing glacial chronologies is an important part of paleoclimate research. Radiocarbon methods can be used in cases where organic matter can be associated with glacial events (Lowe, 1991; Clark et al., 1995). The advent of surface-exposure methods has made dating possible for older events (Fabel and Harbor, 1999); when erosion rates are exceptionally low, surface-exposure methods can date events back to several million years (Nishiizumi et al., 1991). Such dating helps, for example, in determining the sequence of glacial events in a particular area (Gosse et al., 1995).

Surface-exposure dating can be applied to bedrock that has been freshly exhumed by a moving glacier and to moraine boulders that have been excavated by glacial action (Zreda and Phillips, 1995; Steig et al., 1998; Bierman et al., 1999; Marsella et al., 2000; Owen et al., 2002). When dating moraines, replicate samples are usually taken to help in dealing with possible complications (Zreda and Phillips, 1995). For example, erosion of the moraine brings fresh boulders to the surface, giving ages that are too young; some boulders may be reworked from older deposits and appear too old (Hallet and Putkonen, 1994).

Tectonic Chronology

Surface-exposure dating is making an important contribution toward understanding tectonic events over the past few million years. Establishing chronologies for movements along faults allows slip rates to be determined and earthquakes to be dated. Two ways to establish such chronologies involve surface-exposure dating of fault scarps (Zreda and Noller, 1998; Mitchell et al., 2001) and of alluvial sediments that have been affected by fault motion (Brown et al., 1998; Van der Woerd et al., 1998; Zehfuss et al., 2001).

The displacement history of a limestone fault scarp in Israel was studied by using ^{36}Cl by Mitchell et al. (2001). The ^{36}Cl concentration in the exposed footwall as a function of height was determined; this concentration increases with height because the higher positions on the footwall were exposed earlier and have received a larger dose of cosmic rays. Mitchell et al. have discussed methods of inferring the most likely displacement scenario from the ^{36}Cl data and explicitly considered subsurface production rates and the slow exhumation of the sampled surface as faulting progressed.

The Holocene slip rate of the Xidatan segment of the Kunlun fault in Tibet was determined by Van der Woerd et al. (1998). Surface-exposure dating was applied to alluvial sediments that had been offset by fault motion. The dating used both ^{26}Al and ^{10}Be in quartz. By combining the sediment dates with the measured offsets, the authors were able to compute a slip rate.

Volcanic Eruption Dating

Surface-exposure dating of the cooled lava can serve to date the eruption of a volcano. The usual AMS nuclide used is ^{36}Cl, often in conjunction with the stable nuclide ^{3}He. We note that for young basalts with a low K content, the usual method, ^{40}Ar/^{39}Ar, may be ineffective. An example of this application is provided by the work of Zreda et al. (1993), who dated the eruption complex at Lathrop Wells, Nevada, by using ^{36}Cl.

Dating Fluvial, Lacustrine, and Marine Terraces

Surface-exposure dating is relatively straightforward on bedrock, but can be more complicated when dating sediment deposited on a river terrace or a beach. In such an environment, the sediment grains are initially laid down with an ‘‘inherited’’ cosmogenic nuclide concentration derived from prior exposure in the sediment source area (Trull et al., 1995; Anderson et al., 1996). Accurate exposure dating requires determination of the inheritance, which is then subtracted from concentrations at the surface to yield the postdepositional exposure time. The inheritance is usually determined by measuring a depth profile of the cosmogenic nuclide concentration; the inherited and postdepositional components are determined by fitting to the expected profile (e.g., Hancock et al., 1999). This approach works well when the depositional fabric is preserved, but if the sediment or soil has been mixed, then a bulk sample must be taken from the mixed layer as well to determine the total exposure time (Perg et al., 2001).

Erosion Rates

Soil-mantled landscapes seldom have surfaces that are amenable to exposure dating, but cosmogenic nuclides in soils and sediments can be used to infer erosion rates. Exposure histories of individual grains are complicated, as bedrock is weathered and turned to soil, and the soil is then eroded and transported downhill. In this situation, cosmogenic nuclides can be used to infer the residence time of mineral grains within the uppermost bedrock and soil (Lal, 1991), which in turn depends on the erosion rate. The average cosmogenic nuclide concentration in bulk sediment samples can be employed to determine erosion rates of sediment source areas (see, e.g., Brown et al., 1995; Bierman and Steig, 1996; Granger et al., 1996; Clapp et al., 2000, 2001, 2002; Riebe et al., 2000; Schaller et al., 2001). Landscape erosion rates determined in this way are used to measure rates of landscape evolution (e.g., Braun et al., 2001; Granger et al., 2001b), rates of soil formation from bedrock (e.g., Heimsath et al., 2000), and changes in erosion rate through time (Granger et al., 2001a).

Burial Dating of Sediment

The buildup of unstable cosmogenic nuclides in clasts exposed at the surface, and the subsequent decay of these nuclides if the clasts are then buried, has led to a method for dating the burial time of sediments (Lal and Arnold, 1985; Klein et al., 1986; Lal, 1991; Granger et al., 1997; Granger and Muzikar, 2001). Clasts initially on the surface are exposed to cosmic rays and so accumulate an inventory of in situ cosmogenic nuclides. The sediment is then buried, for example, by being covered with a thick layer of overburden or by being washed into a cave. After the burial event, the sediment is shielded from cosmic rays, and so the concentration of each unstable cosmogenic nuclide decreases with time. This decrease allows one to determine how much time has passed since the burial.

Because the initial concentrations of the cosmogenic nuclides at burial are in general not known, the common approach is to model the preburial history in some way. This approach introduces at least one unknown parameter, such as the preburial exposure time. So, there are at least two unknowns to be determined: the burial time and any parameters that enter into the preburial history model. Thus, at least two different cosmogenic nuclides must be used in burial dating.

One common application of this technique uses ^{26}Al and ^{10}Be in quartz and has an effective range of ~100 k.y. to 5 m.y. (Granger and Muzikar, 2001). The lower limit of 100 k.y. is set by the fact that at present levels of accuracy, the uncertainty in a burial age is ~100 k.y. The upper limit is set by several considerations. After the burial time reaches four or five half-lives, the concentrations of ^{10}Be and ^{26}Al become small and more difficult to measure accurately. Also, when the concentrations of the nuclides produced in the preburial stage have become very low, even a small amount of postburial production can lead to a serious loss in precision.

Burial dating with just one nuclide is in principle possible by using ^{14}C produced in situ in quartz. Because ^{14}C has a short half-life, one may assume that at the time the quartz was buried, the ^{14}C was at saturation. Then the concentration of the ^{14}C in the buried quartz directly reveals the burial time. This method has an upper age limit of ca. 30 ka.

The dating of sediment burial events has many potential applications in geology. So far, one important application has been the determination of river incision rates (Granger et al., 2001a). If an incising river leaves sediment deposited in a cave, burial dating of this sediment can reveal when the river was at the cave's level. This method also has a potential application in dating strata at archaeological sites.

Climate and Magnetic Field History

The ice sheets of Greenland, Antarctica, and other regions of the world contain layers of ice deposited over times ranging back into the Pleistocene and are an invaluable source of information for studying climate variation over the past several hundred thousand years (Lowe and Walker, 1997).

The study of ^{10}Be and ^{36}Cl in ice cores is an active and still developing field (Baumgartner et al., 1997a; Finkel and Nishiizumi, 1997; Wagner et al., 2001). After being produced in the atmosphere, these nuclides are deposited on the ice either through precipitation or dry deposition. Their concentration in the ice varies with depth and is a function of several factors (Beer et al., 2002): (1) Variations in the Earth's magnetic field, and in the solar cycle, will affect the atmospheric production rate. (2) The annual precipitation rate will affect the nuclide concentrations of the ice because the more ice formed in a given year, the lower the concentrations of the ^{10}Be and the ^{36}Cl. (3) Atmospheric circulation patterns can have an effect because cosmogenic nuclides produced in the atmosphere at a given location can be deposited in a quite different location. At the polar latitudes of many ice sheets, variations in the Earth's magnetic field will have little effect on the local atmospheric nuclide production; thus, variations in the nuclide concentration in the ice due to variations in the Earth's field are largely due to cosmogenic nuclides produced at lower latitudes. (4) One complicating factor is that terrestrial dust particles containing ^{10}Be produced long ago can be blown into polar regions and deposited in the ice (see Baumgartner et al., 1997b). (5) The concentrations of both nuclides will be reduced by nuclear decay. If the oldest ice was deposited at 400 ka, for example, the ^{10}Be concentration will be reduced to 83% of its initial value.

Thus, the ^{10}Be and ^{36}Cl concentrations in the ice cores contain information concerning the history of the Earth's magnetic field, the solar cycle, and the Earth's climate regimes. A key challenge in this research is unraveling all the various effects that are embodied in the nuclide signal. One technical aspect of this research is that a careful analysis of a long ice core usually requires many AMS measurements; progress in this field is therefore related to increased efficiency of AMS measurements of ^{10}Be and ^{36}Cl (Fifield, 2000).

Cosmogenic nuclides can also be studied in sediment cores. Beneath certain lakes (see, e.g., Horiuchi et al., 2000), at the bottom of certain ocean regions (see, e.g., Aldahan and Possnert, 2000; Frank, 2000), or in regions of loess accumulation (see, e.g., Gu et al., 1996) are layers of sediment that have been accumulating for thousands of years without interruption. ^{10}Be is the cosmogenic nuclide most commonly studied in such sediment.

Radiocarbon Dating

This is a vast subject (see, e.g., Wagner, 1998); here we will briefly discuss the particular points relevant to AMS (see, e.g., Tuniz et al., 1998). With AMS, one directly measures the ^{14}C/^{13}C (or ^{14}C/^{12}C) ratio, and through a series of calculations this is turned into a date (for a clear explanation of these calculations, see McNichol et al., 2001).

The ability to date small samples, containing 1.0–0.1 mg or even less of carbon, has important implications for geologic research. Rather than date all the carbon in a given sample, the carbon from a particular component can be dated. For example, in a sediment core, one can date just the pollen, rather than all the organic matter (Brown et al., 1989). Or one can date a particular chemical component, such as the lignin (McNichol et al., 2000).

The main difficulty in dating samples that are too old (i.e., older than 40 ka) or too small (i.e., <0.1 mg of carbon) is contamination. AMS can in principle count the ^{14}C atoms in a very old sample; the problem is that during the chemical preparation a small amount of contaminating carbon is inevitably added to the sample. If the sample is too old or too small, the number of ^{14}C atoms due to the contaminant will be comparable to the number of ^{14}C atoms in the original sample. So, even if AMS can accurately count the ^{14}C atoms, the date will be incorrect.

This issue is being actively addressed in two ways (Kirner et al., 1995; Brown and Southon, 1997): (1) The amount of contaminating carbon added during processing can be determined, and its effects subtracted out. (2) The chemical preparation can be improved to reduce the amount of contaminating carbon added. At present, the oldest age that can be accurately determined by AMS radiocarbon dating is ca. 50 ka. The lower limit of sample size is on the order of tens of micrograms of carbon. So far, the contamination problem has frustrated hopes that AMS could push the radiocarbon age limit back to well beyond 50 ka (Fifield, 2000).

Subduction-Zone Sediments

When an oceanic plate is subducted and island-arc volcanoes are formed, some of the sediment carried by the subducting plate is incorporated into the resulting magma. Such sediment contains ^{10}Be (produced in the atmosphere and adhering to the sediment) that can be used to measure the fraction of sediment in the magma and the time it takes for the sediment to be recycled into lava (Tera et al., 1986; Monaghan et al., 1988; Morris et al., 1993). Other work has applied ^{129}I, which resides in sediment fluids, to this same problem (Snyder and Fehn, 2000).

Magma originating from the mantle, or from the sub-sedimentary part of the subducting plate itself, will contain very little ^{10}Be, owing to its nuclear decay. Thus, if this nuclide can be detected in the magma, it very likely comes from the sediment covering the plate. Because ^{10}Be has a half-life of 1.5 m.y., its presence in the magma also means that the time for the sediment to be transported through the subduction zone and finally emerge in the magma cannot be longer than 5–10 m.y.

^{10}Be can also be useful in complex tectonic situations in which it is not clear whether volcanoes are due to subduction. As Morris et al. (1993) have pointed out, the presence of this nuclide in the magma would be strong evidence for a subduction mechanism.

Hydrology

AMS nuclides have been widely applied in studies of groundwater, lakes, and oceans (Elmore et al., 1979; Fontes and Andrews, 1994; Tuniz et al., 1998). Before the advent of AMS, hydrologic studies using ^{14}C with decay counting were done; the smaller sample size needed by AMS means that much smaller water volumes can be utilized. The main nuclides used are ^{14}C, ^{36}Cl, and ^{129}I; a knowledge of the chemical behavior of each of these elements is crucial in understanding the basis for this work. In many ways, the simplest behavior is that of ^{36}Cl. Chlorine remains dissolved in water as the Cl^- ion over a wide range of conditions. Iodine also tends to stay in water as the I^- ion, but it has a competing tendency to be incorporated in organic matter in several oxidation states. Carbon exists in water in many different forms—as dissolved CO_2, as $CO_3^{2-}$and HCO_3^- ions, and in organic molecules.

As discussed in previous sections, the three nuclides used in hydrology are produced in the atmosphere by cosmic rays. Other important production mechanisms sometimes come into play in hydrology, however. The atmospheric nuclear bomb tests led to a great increase in the atmospheric concentrations of these three nuclides (Synal et al., 1990; Fontes and Andrews, 1994; Levin and Kromer, 1997; Tuniz et al., 1998). Since the end of the atmospheric tests, these ''bomb-pulse'' nuclides have propagated into different components of the hydrosphere and are proving useful as tracers (Cecil and Vogt, 1997; Milton et al., 1997). Intrinsic radioactivity in the lithosphere also can lead to production of these three nuclides (Andrews et al., 1989; Tuniz et al., 1998). Alpha particles emitted from U, Th, and their decay products can produce slow neutrons when they interact with nuclei; these neutrons can be captured by ^{35}Cl to produce ^{36}Cl. The fission of ^{238}U can produce ^{129}I as a by-product.

One hydrologic application of AMS nuclides is the dating of water. Here, the date established is the time the water was last in contact with the atmosphere; the half-life of the nuclide determines the age range that can be covered (Tuniz et al., 1998). When in contact with the atmosphere, water acquires a certain concentration of the unstable cosmogenic isotope from the atmosphere; the water thus has an initial ratio of $^{129}I/^{127}I$, $^{14}C/^{12}C$, or $^{36}Cl/Cl$. As the water moves away from its contact with the atmosphere, this ratio decreases owing to radioactive decay of the unstable isotope. Measurement of the ratio reveals how much time has passed since the water was in contact with the atmosphere (Torgersen et al., 1991).

A related application involves tracing cosmogenic nuclides through the environment in order to quantify water flow rates. The study by Phillips et al. (1988), for example, used the bomb-pulse peak in ^{36}Cl to measure recharge rates for a sandy soil in New Mexico. Balderer and Synal (1996) used ^{36}Cl to characterize groundwater circulation in the North Anatolian fault zone.

NEW DEVELOPMENTS IN AMS FOR EARTH SCIENCE APPLICATIONS

Important developments are arising in the AMS technique itself and in the application of AMS to geologic research (for a related discussion, see Fifield, 2000). One important advance is the development of a new generation of AMS systems in which much lower voltages, on the order of 0.5 MV, are used (Synal et al., 2000). So far, these machines have been used only for ^{14}C measurements, but they could potentially be applied to ^{26}Al and ^{129}I. These lower-voltage machines are much less expensive to buy and use, are simpler to operate, and require less space.

As we have explained, the high voltages used in the usual machines have several purposes: (1) In order to strip several electrons from unwanted molecules and hence cause their breakup by the stripper gas or foil, the molecules are accelerated to high velocities; and (2) distinguishing interfering isobars in the final detector requires that the ions be accelerated to high energies. In some cases a lower voltage can work: (1) Synal et al. (2000) have shown that even though only one electron is stripped off the molecules, they can still be broken up by collisions if the thickness of the stripper gas is large enough, and (2) certain nuclides, ^{14}C, ^{26}Al, and ^{129}I, do not have isobars that form negative ions. For nuclides that do have isobars that form negative ions (^{10}Be and ^{36}Cl), it will be more difficult to develop low-voltage techniques.

Another advance lies in the development of new AMS nuclides for geologic research (Tuniz et al., 1998; Fifield, 1999). Nuclides that may play a role in the future include ^{32}Si, ^{39}Ar, and ^{41}Ca. The nuclide ^{32}Si, which has a half-life of ~140 yr, is being used as a tracer in a variety of geologic settings, and naturally occurring amounts can be measured with AMS (see, e.g., Treacy et al., 2000; Morgenstern et al., 2000). The nuclide ^{39}Ar, which has a half-life of ~270 yr, is finding use as a tracer in oceanography (see Broecker and Peng, 2000). So far, however, decay counting has been used for this nuclide because AMS is quite difficult at the naturally occurring concentrations (Kutschera et al., 1994). Finally, ^{41}Ca is being measured with AMS for biomedical applications, where its concentration can be made much higher than in naturally occurring samples (Jackson et al., 2000). AMS has been used to detect ^{41}Ca in natural samples (Kutschera, 1990; Fink et al., 1990; Finkel and Suter, 1993), but the measurements are difficult. This nuclide is produced in rocks by cosmic rays (mainly by neutron capture on ^{40}Ca) and so has potential geologic applications.

The ability of AMS to measure the very small natural concentrations of certain nuclides has opened new possibilities for viewing geologic events and processes. This AMS revolution is having an impact in many diverse fields, as we hope to have shown in this paper. New applications are appearing, and we expect this trend to continue as geologists become more familiar with AMS.

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