data is available from J.W.V.

22. Conventional oxygen isotope analyses were made of 10-mg mineral separates concentrated from a cube, 4 cm on a side (sample 90LP9a) cut from the larger block. Ion microprobe analyses were made on a thin section 2.5 cm in diameter that was cut along one face of the cube. Small pieces of calcite were removed from each of the corners of the cube; the remainder was dissolved in dilute HCl. Magnetites were then handpicked from the residue. Calcites were analyzed by conventional reaction with concentrated (density = 1.92) phosphoric acid. One large magnetite grain was powdered and analyzed by reaction with BrF$_5$ in conventional nickel reaction vessels that were externally heated by electric resistance furnaces. Other magnetites were reacted with BrF$_5$, either as powder or as intact single crystals, with a CO$_2$-laser heat source (33). The conventional and laser analyses yield similar accuracy and precision (Table 1), but the laser permits significantly smaller sample sizes.

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The ^{18}O/^{16}O Ratio of 2-Billion-Year-Old Seawater Inferred from Ancient Oceanic Crust

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An oxygen isotope profile of the 2-billion-year-old Purtuniq ophiolite overlaps with similar profiles of younger ophiolites and the modern oceanic crust. This overlap implies (i) that there was a similar style of seawater–ocean crust interaction during the past 2 billion years; (ii) that the oxygen isotope composition of early Proterozoic seawater was similar to the modern value; (iii) that early Proterozoic sea-floor spreading rates were similar to, or greater than, average modern rates; and (iv) that early Proterozoic carbonate rocks and cherts with low ^{18}O/^{16}O ratios do not reflect global-scale ^{18}O depletion of early Proterozoic oceans.

A long-standing controversy concerns temporal variations in the oxygen isotope composition of Precambrian seawater. It has been suggested that the δ^{18}O (1) record of carbonates (2–4) and cherts (5) over the past 2 billion years reflects variations in the isotopic composition of the seawater in which these sediments were formed. Others argue that pervasive ^{18}O exchange between seawater and the oceanic crust caused the δ^{18}O value of seawater to remain constant over time (6, 7). Satisfactory resolution of such conflicting views is important in that the oxygen isotope composition of Precambrian seawater bears on the nature of Precambrian plate tectonics and climate.

In this report we show that the imprint of early Proterozoic seawater is preserved in the δ^{18}O profile of the 1.998-billion-year-old Purtuniq ophiolite, situated in the Cape Smith Belt of northern Quebec, Canada (8–10). The controversy surrounding the true δ^{18}O value of early Proterozoic seawater is addressed through a comparison of the isotopic imprint of 2-billion-year-old seawater determined for the ophiolite with that in contemporaneous carbonate rocks of the ~1.9-billion-year-old Rocknest dolostone, situated in the Wopmay orogen, Northwest Territories, Canada (3, 4). The opportunity to

analyze samples from the lower levels of early Proterozoic oceanic crust distinguishes this work from most earlier studies that have utilized less desirable submarine pillow basalts collected from Precambrian greenstone terranes (11, 12).

From their studies of the Rocknest dolostone, Burdett et al. (3) and Veizer et al. (4) concluded that the δ^{18}O value of early Proterozoic calcite was nearly 10 per mil lower than that of modern marine calcites, which are ~30 per mil (relative to SMOW). On the basis of the excellent preservation of original marine fabrics observed in thin section, Burdett et al. argued that dolomitization of the primary carbonates did not completely overprint the original seawater signature. Using the isotopic composition of the most ^{13}C- and ^{18}O-enriched dolomitized ooid of 22.9 per mil and an assumed dolomite-calcite fractionation of +4 per mil, Burdett et al. calculated a δ^{18}O value of 18.9 per mil for the original Rocknest calcite. If equatorial seawater temperatures in the early Proterozoic were similar to today's, Burdett et al. (3) suggested, as have others (4, 5), that lower δ^{18}O values for early Proterozoic sediments may reflect a nearly 10 per mil ^{18}O depletion in early Proterozoic seawater.

Although widely accepted, the above proposal conflicts with predictions that the average δ^{18}O value of the global ocean

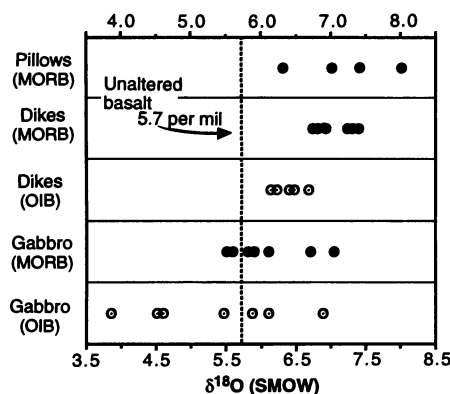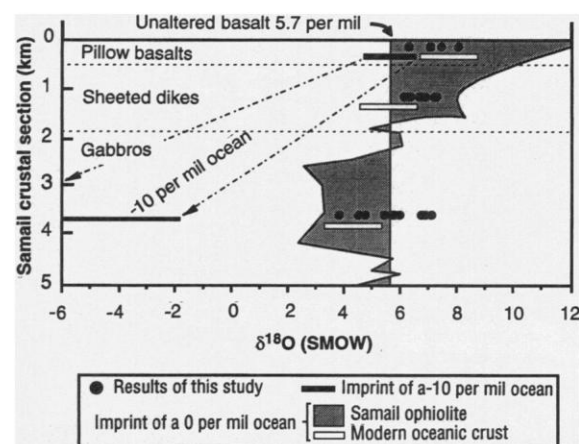Department of Geology, University of Alberta, Edmonton, Alberta, Canada T6G 2E3.

Fig. 1. The $\delta^{18}O$ profile of submarine pillows, sheeted dikes, and gabbros of the Purtuniq ophiolite presented by suite, (●) MORB and (○) OIB. The MORB and OIB suite designations refer to interpretations in (18). Analytical uncertainties are about the size of the symbols (36).

Fig. 2. Composite $\delta^{18}O$ profile of the Purtuniq ophiolite compared to the $\delta^{18}O$ profile of the Samail ophiolite (7), which reflects the characteristic imprint of an ~0 per mil ocean. The range of isotopic values for the modern oceanic crust is from (6, 24). The hypothetical imprint of a −10 per mil ocean on the oceanic crust is also displayed (37). The shape of the observed $\delta^{18}O$ profile and the magnitude of the $\delta^{18}O$ values are entirely consistent with an early Proterozoic seawater isotopic composition of 0 ± 2 per mil (SMOW).

remained near 0 ± 2 per mil for most of Earth history (6, 11). Presently, the $\delta^{18}O$ value of seawater is largely determined by its interactions with the oceanic crust. Seawater, which penetrates the oceanic crust at ridges, reacts with newly formed crust ($\delta^{18}O = 5.7 ± 0.3$ per mil) at both low and high temperatures. In upper crustal levels, low-temperature (10° to 150°C) alteration decreases the $\delta^{18}O$ value of seawater but increases the $\delta^{18}O$ value of the rock (13). Deeper in the crustal section, at temperatures of ~350°C, the polarity of exchange reverses and the $\delta^{18}O$ value of seawater is increased at the expense of the rock (14). The ^{18}O enrichments balance the ^{18}O depletions, in effect buffering the $\delta^{18}O$ value of seawater and imprinting the crustal section with a characteristic $\delta^{18}O$ profile (6, 7). The shape of this imposed profile and the magnitude of the $\delta^{18}O$ values reflect the isotopic composition of the circulating seawater. If early Proterozoic seawater was ^{18}O-depleted relative to modern seawater, then the 2-billion-year-old ophiolite would likewise be ^{18}O-depleted compared to modern oceanic crust.

The 2-billion-year-old Purtuniq ophiolite is part of a package of mafic crustal rocks that were juxtaposed during convergence of early Proterozoic continental and oceanic lithospheric plates (15, 16). Three tectonic settings are inferred for rocks of the belt: a rifted continental margin, an oceanic spreading center, and an island arc–related subduction center. The ophiolitic portion preserves all the requisite components of the modern oceanic crust in a series of thrust-bounded sheets that are metamorphosed predominantly to upper greenschist facies (17). This is a slightly higher grade than the lower greenschist conditions prevalent during seawater-induced alteration of the deep oceanic crust

today. Samples analyzed in this study contain hornblende or actinolite + plagioclase + quartz, ± chlorite, ± epidote, ± calcite, and ± sphene (8). The belt is argued to be direct evidence for the operation of Phanerozoic-style plate tectonics 2 billion years ago (15).

Field and chemical evidence indicates that two magma sources formed the Purtuniq crustal section (9, 18): an older tholeiitic suite, chemically similar to modern mid-ocean ridge basalts (MORB) with $\epsilon_{Nd}(t) = 4.0$ to 5.3 and a younger suite chemically similar to ocean island type basalts (OIB) with $\epsilon_{Nd}(t) = 2.5$ to 3.6, where $t = 2$ billion years (18, 19). Most thrust sheets are characterized by rocks of either MORB or OIB affinity with the exception of the sheeted dikes and one example of cross-cutting mafic cumulates (20). The composite crustal section (MORB + OIB) is about 9 km thick with roughly equal contributions from each source (21).

From the pattern observed in Fig. 1, it is evident that the Purtuniq ophiolite displays the characteristic isotopic signature that results from seawater–ocean crust interaction, specifically, ^{18}O enrichment in the pillow and dike sections in contrast to ^{18}O depletion in the gabbros. The shape of the profile is evidence that seawater, initially cooler and heating up with depth, penetrated and altered the Purtuniq crustal section. The ~0.7 per mil offset between MORB and OIB suites (Fig. 1) probably reflects the superposition of the effects of separate hydrothermal circulation cells associated with multiple intrusions. Preservation of the $\delta^{18}O$ differences between rock units is evidence against regional-scale isotopic homogenization during later metamorphism. Distinct isotopic signatures for adjacent MORB and OIB dike samples rule out significant isotopic resetting of the whole rock.

The $\delta^{18}O$ value of 2-billion-year-old seawater is best inferred from the gabbro sec-

tion of the ophiolite. Low-temperature alteration products of pillow basalts, as with the low-temperature formation and subsequent diagenesis of low-temperature carbonate rocks and cherts, involve large mineral-water $\delta^{18}O$ fractionations (for example, $\Delta_{calcite-water} \approx +30$ per mil, at 20°C). Consequently, the final $\delta^{18}O$ value of the rock is sensitive to the temperature as well as the extent of alteration and renders sediments and pillow basalts poor monitors of potential changes in the $\delta^{18}O$ value of the oceans. At higher temperatures, isotopic fractionations between interacting water and rock are smaller (for example, $\Delta_{basalt-water} \approx +6$ per mil, at 350°C), and uncertainty over the exact temperature of in situ hydrothermal alteration is less important in determining the final $\delta^{18}O$ value of the rock. Therefore, if seawater were genuinely ^{18}O-depleted to the extreme values suggested (3–5), the gabbros would be significantly depleted of ^{18}O, irrespective of any uncertainty in temperature.

The $\delta^{18}O$ profile of the Purtuniq ophiolite (Fig. 2) overlaps with similar profiles reported for younger ophiolites (22, 23) and the modern oceanic crust (13, 14, 24), consistent with an isotopic composition for 2-billion-year-old seawater near the modern value of 0 per mil. Had the oceans been at −10 per mil, gabbros with $\delta^{18}O$ values as low as −8 per mil would have been evident. No such ^{18}O-depleted rocks were observed.

The Purtuniq ophiolite and Rocknest dolostone results may be reconciled through (i) higher equatorial seawater temperatures in the early Proterozoic, (ii) diagenetic overprints, and (iii) decoupling of the oxygen isotope systematics between epeiric seas and the deep global ocean. Although Burdett et al. (3) argued that they were able to see through the diagenetic overprint to the imprint of the original seawater, they conceded that their data are compatible with either ^{18}O-depleted oceans or higher equatorial seawater temperatures 2 billion years

Table 1. Summary of $\delta^{18}O$ results.

Sample	Type	$\delta^{18}O$ (SMOW)*
	Basalts	
S-198	MORB	7.0
S-199	MORB	7.4
S-201	MORB	6.3
D-247a	MORB	7.4
B-367	MORB	8.0
	Dikes	
C-5	MORB	7.4
C-11	MORB	6.9
B-1	MORB	6.8
B-2	MORB	7.3
B-5	MORB	6.7
E-7	MORB	7.3
D-3	OIB	6.4
E-1	OIB	6.2
E-4	OIB	6.1
E-5	OIB	6.6
E-6	OIB	6.3
	Gabbros	
B-407	MORB	5.8
D-143a	MORB	5.5
D-144a	MORB	5.6
D-248f	MORB	5.9
S-212a	MORB	7.0
S-213	MORB	6.7
L-181p	MORB	6.1
L-181m	MORB	5.8
S-315	OIB	4.5
S-316	OIB	3.9
S-317	OIB	5.5
S-326	OIB	6.9
S-345a	OIB	5.9
B-369a	OIB	5.5
B-374	OIB	4.6
B-414a	OIB	6.1

*Isotopic values reported in the standard delta (δ) notation as per mil deviation from SMOW (1).

ago; they did not consider the third alternative. Using the criteria outlined in (3), we calculate a temperature of 75°C for precipitation of the original Rocknest calcite if seawater were −1 per mil (25). The Rocknest temperature falls within the range of temperatures calculated for various phases of the ~2-billion-year-old Gunflint banded iron formation. For example, Winter and Knauth reported (26) an average chert $\delta^{18}O$ value of 23.2 ± 0.08 per mil (n = 82), which, depending on the silicawater fractionation used, corresponds to temperatures between 55° and 90°C. From the same formation, Karhu and Epstein (27) reported paleotemperatures from chert, phosphate, and chert-phosphate pairs of 84°, 95°, and 66°C, respectively, the last determination being independent of the $\delta^{18}O$ value for seawater.

The inferred paleotemperatures are often criticized as unreasonably high in that sedimentary evidence for early Proterozoic glaciations (28) does not appear to corroborate the idea of hot Precambrian oceans. However, Knauth and Lowe (29) pointed out that, on geological time scales, glaciations would be short-lived phenomena and

would not conflict with an overall hot climate. Furthermore, because it is well established that nearly all post-depositional alteration phenomena lead to ^{18}O depletion of the original sediment, the lowest calculated temperatures are the most reliable. Alternatively, it may be argued that the original seawater imprint is often not preserved in Precambrian sediments. Diagenesis transforms the primary metastable precipitates of aragonite and high-magnesium calcite, hydrous silica, and carbonate-hydroxyapatite to the diagenetically stable neomorphs; dolomite, quartz, and apatite. Because most isotopic analyses of Precambrian sediments are performed on secondary phases, the high calculated temperatures may reflect the fluid and temperature conditions of burial diagenesis, including dolomitization, rather than those of the original ocean.

Another potentially important source of uncertainty in carbonate- and chert-based paleotemperature estimates is the assumption that the $\delta^{18}O$ value of the water from which Precambrian sediments precipitated was isotopically identical to that of the deep ocean. Many Precambrian carbonate rocks and cherts probably formed in shallow epeiric or restricted continental-shelf seas that are inherently more susceptible to the influx and entrainment of low-$\delta^{18}O$ meteoric waters. For example, although the $\delta^{18}O$ value of the Cretaceous ocean was 0 per mil, waters of the epicontinental Greenhorn Sea, Cretaceous Western Interior Seaway, were depleted in ^{18}O by as much as 4 per mil (30). If the majority of Precambrian sediments formed from ^{18}O-depleted seawater in marginal marine settings, then the temperatures calculated need not be as high as the observed ^{18}O depletion; that is, for every 1 per mil decrease in the $\delta^{18}O$ value of a marginal marine sea, the inferred paleotemperature decreases by about 4°C.

Although the above considerations hamper a confident interpretation of the $\delta^{18}O$ variations observed in Precambrian sediments, the isotopic data garnered from ophiolites is far less equivocal on one important issue, that of the $\delta^{18}O$ value of seawater. The imprint of 2-billion-year-old seawater preserved within the Purtuniq ophiolite clearly shows that the isotopic composition of early Proterozoic seawater was similar to that of modern seawater. In accordance with model predictions (6), the $\delta^{18}O$ value of early Proterozoic seawater may be constrained to the range $-2 \leq \delta^{18}O \leq +2$. Because the $\delta^{18}O$ value of the oceans reflects the cycling of seawater through the oceanic crust, a 0 per mil ocean in the early Proterozoic is consistent with widespread sea-floor spreading operating in that time. Modern studies have shown that

sea-floor spreading is the driving force behind the production and maintenance of a 0 per mil ocean (6, 11). Beneath Earth's extensive mid-ocean ridge spreading centers, hot upwelling magmas drive convective circulation of seawater through the oceanic crust and promote pervasive ^{18}O exchange between water and rock at high temperatures. By removing older altered crust, sea-floor spreading provides a mechanism by which new oceanic crust is continually brought into contact with deep convecting seawater. The proposed constancy of the isotopic composition of seawater over time (6) reflects the constancy of the grander process of production, alteration, and destruction of the oceanic crust.

This fundamental relation between the isotopic composition of seawater and sea-floor spreading may extend to even older Archean oceans. Archean carbonate rocks (31) and cherts (5, 29) yield ^{18}O-depleted values similar to, or less than, those of early Proterozoic sediments. In contrast, isotopic studies of proposed Archean oceanic crust—for example, the Barberton Greenstone Belt (12, 32) and the Pilbara Block (11)—and of Archean hydrothermal ore deposits (33, 34) suggest that the $\delta^{18}O$ value of the oceans has remained close to 0 per mil for the past 3.5 billion years. Isotope budget modeling of the modern oceans (6, 11, 35) suggests that sea-floor spreading rates that are at least ~50% of modern average rates are rapid enough to control the $\delta^{18}O$ value of seawater. Because the isotopic composition of seawater reflects, in a fundamental way, plate tectonic processes such as mantle convection and sea-floor spreading, global-scale ^{18}O-depleted Precambrian oceans would be possible only if the modern plate tectonic paradigm were considered invalid for most of Earth history.

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37. Conservative estimates for the $\delta^{18}O$ values of pillow basalts reflect modern observations of recent sea-floor material and are based on mixtures of 10% clay ($\Delta_{clay-water} \approx +25$ per mil, at temperatures typical of modern seawater) and 90% basalt (+5.7 per mil). The basalt $\delta^{18}O$ whole-rock values change by only ~ 1 per mil, even though the $\delta^{18}O$ value of seawater changes by 10 per mil. The $\delta^{18}O$ value of a whole-rock

Released Form of CNTF Receptor α Component as a Soluble Mediator of CNTF Responses

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The α component of the receptor for ciliary neurotrophic factor (CNTF) differs from other known growth factor receptors in that it is anchored to cell membranes by a glycosyl-phosphatidylinositol linkage. One possible function of this type of linkage is to allow for the regulated release of this receptor component. Cell lines not normally responsive to CNTF responded to treatment with a combination of CNTF and a soluble form of the CNTF α receptor component. These findings not only demonstrate that the CNTF receptor α chain is a required component of the functional CNTF receptor complex but also reveal that it can function in soluble form as part of a heterodimeric ligand. Potential physiological roles for the soluble CNTF receptor are suggested by its presence in cerebrospinal fluid and by its release from skeletal muscle in response to peripheral nerve injury.

Originally, CNTF was described as an activity that supports the survival of ciliary ganglion neurons in vitro (1), but it is now known to promote the survival or differentiation (or both) of a wide assortment of cells within the nervous system (2). The molecular cloning of a CNTF-binding protein (here denoted CNTFRα) revealed that it was similar to one of the two components of the interleukin-6 (IL-6) receptor (here denoted IL-6Rα) (3, 4). This similarity suggested that, like the IL-6 receptor, which has a signal-transducing component called gp130 (5, 6), the CNTF receptor might also require additional components to effect signal transduction (4).

The factors CNTF and IL-6 are distantly

related members of a family that includes leukemia inhibitory factor (LIF) [also known as cholinergic differentiation factor (CDF) (7, 8)]; all of these factors utilize multicomponent receptors that share signal-transducing subunits (9, 10). In particular, all of these receptor complexes include gp130 (10). Both CNTF and LIF (but not IL-6) may also share an additional receptor subunit that is related to gp130 (here referred to as LIFRβ) (10), which was initially identified as a LIF-binding protein (11). It has been suggested that gp130 and LIFRβ together form a functional LIF receptor complex (9, 10) and that addition of the CNTFRα chain to this complex is sufficient to convert a functional LIF receptor into a functional CNTF receptor (10). Whereas the broad distributions of both gp130 and LIFRβ presumably allow for the widespread actions of LIF, the limited distribution of CNTFRα (predominantly to the nervous system) may restrict CNTF to its neuronal actions (12).

The CNTF receptor is unlike all other

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