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The Age and Origin of the Little Diomede Island Upland Surface

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ABSTRACT. Geomorphology and projected uplift rates indicate that the upland surface of Little Diomede Island may represent a high sea level stand that occurred 2.6 million years ago in the Bering Strait. The 350–363 m upland surface of the island could be correlative with the York terrace, an uplifted marine terrace previously recognized on the southern flanks of the York Mountains, Seward Peninsula. The modern surface of Little Diomede Island is composed of a cryoplanation terrace enclosing a central blockfield and rimmed with tors. Beryllium-10 cosmogenic isotope analysis of two tors and three outcrops from the upper surface indicate the island has been under the influence of a subaerial periglacial environment at least for the last 36 000 years (MIS 3) and probably for 254 000 (MIS 7/8). Unequivocal evidence does not exist to support glaciation of Little Diomede Island.

Key words: Little Diomede Island, Bering Strait, York terrace, cosmogenic isotope dating, beryllium-10

RÉSUMÉ. La géomorphologie et les taux d'exhaussement obtenus par extrapolation révèlent que la surface de haute terre de l'île de Petite Diomède pourrait représenter un relief ayant existé dans le contexte d'un niveau de mer élevé qui avait cours il y a 2,6 millions d'années dans le détroit de Béring. La surface de haute terre de l'île, atteignant de 350 à 363 m, pourrait être en corrélation avec la terrasse de York, terrasse marine surélevée, découverte antérieurement sur les flancs méridionaux des monts York situés dans la péninsule Seward. La surface actuelle de l'île de Petite Diomède se compose d'une terrasse de cryoplanation entourant un champ central de blocs rocheux et circonscrite par des tors. L'analyse isotopique cosmogonique au ^{10}béryllium de deux tors et de trois affleurements de la surface la plus haute révèle que l'île a subi l'influence d'un environnement périglaciaire subaérien pendant au moins les 36 000 dernières années (3^{e} étage isotopique marin) et probablement 254 000 ans (7^{e}/8^{e} étage isotopique marin). On ne possède pas de preuve non équivoque d'une glaciation de l'île de Petite Diomède.

Mots clés: île de Petite Diomède, détroit de Béring, terrasse de York, datation cosmogonique aux isotopes, ^{10}béryllium

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INTRODUCTION

Although it is widely accepted that the Bering Strait repeatedly served as the central part of the Bering Land Bridge connecting Asia and North America (Hopkins, 1967), the Pleistocene history of the Diomede Islands, situated at the narrowest part of the Bering Strait, has remained unstudied. Little Diomede Island (65˚45′N, 168˚56′W) is located 43 km west of Cape Prince of Wales, Alaska, and 4 km east of Big Diomede Island, Russia (Fig. 1). At numerous times in the past, sea level has lowered and the Bering Strait has become emergent, forming a land bridge between the two continents. This research details the evolution of a remnant of the central former land bridge (the Little Diomede Island upland surface). The main objective is to answer two questions: 1) Is the upland surface of Little Diomede Island marine, glacial, or periglacial in origin? and 2) What is the age of this surface?

STUDY AREA

Little Diomede Island is 3 km^2 and rises steeply on all sides from sea level to an undulating plain at 350–363 m (Fig. 2). The village of Inalik (pop. 150) is located on the west side of the island. Historical records indicate that the region is currently submerging. Possibly as late as 135 years ago, a sand spit connected the Little and Big Diomede Islands, with an ephemeral river separating them (Jenness, 1929; C. Ahkinga, pers. comm. 1997).

The regional climate is characterized as subarctic and dominated by extratropical cyclones and arctic anticyclones in winter (Sharma, 1974). The Aleutian Low predominates, causing frequent storms, wind, fog, and precipitation. Because of the cyclonic circulation, the eastern half of the Bering Sea is warmer than the western half; however, the deep southern and western parts of the Bering Sea are generally kept ice-free by

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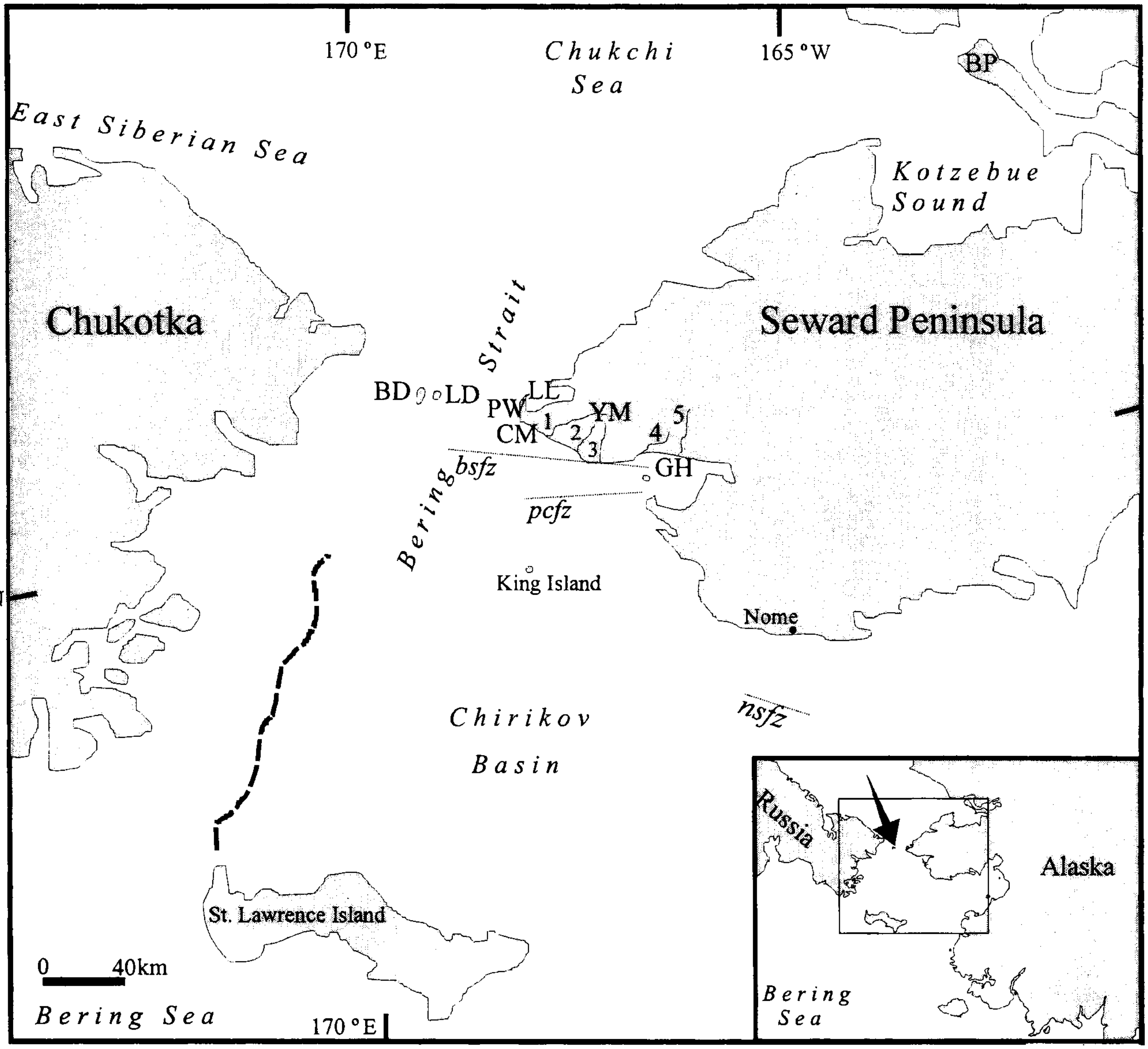

FIG. 1. The Bering Strait region. Arrow on inset shows location of the Diomede Islands and box encloses the enlarged area. Place names are: LD = Little Diomede Island, BD = Big Diomede Island, PW = Cape Prince of Wales, CM = Cape Mountain (site of the York terrace), LL = Lopp Lagoon, YM = York Mountains, GH = Grantley Harbor, BP = Baldwin Peninsula, 1 = Kanauguk River, 2 = King River, 3 = Lost River, 4 = Skull Creek, 5 = California River. Dashed line north of St. Lawrence Island is the offshore linear ridge interpreted by Grim and McManus (1970) and D.M. Hopkins (pers. comm. 1997) to be a moraine. Thin black lines indicate fault zones: *bsfz* = Bering Strait Fault Zone, *pcfz* = Port Clarence Fault Zone, *nsfz* = Norton Sound Fault Zone (from Plafker et al., 1993).

encroachment of warm Pacific water (Sharma, 1974). The maximum sea ice cover in the Bering Strait is in late winter-early spring, while the Strait is generally ice-free from July to September (Grebmeier et al., 1995). Temperatures on Little Diomede Island average 5 to 10˚C in summer and -23 to -14˚C in winter. Winds are predominantly from the north and average 8 m/s, with gusts up to 26–36 m/s (Alaska Department of Community & Economic Development, 1998). In July 1997, snow banks were present on north-facing slopes of Little Diomede Island, as well as on the east-facing slopes of Big Diomede Island. The north-, south-, and west-facing slopes of Big Diomede Island could not be seen or investigated; however, it is possible that late-lying snowpacks exist on these slopes of Big Diomede Island also. The island lies within a zone of discontinuous permafrost (Brown et al., 1997).

The first geologic mapping of the island was by Sainsbury (1972), who determined that it consisted of medium- to coarse-grained, unfoliated biotite hornblende granite of late Cretaceous age. Barker et al. (1994) mapped the island as late Cretaceous granite (97–66 million years old). The predominant granite on the island contains feldspar crystals up to 5 cm long in a coarse-grained matrix of quartz, biotite, and hornblende. The granite has been named the Diomede pluton and is of similar age and lithology as other plutons of the York terrane (Big Diomede Island, King Island) in the Bering Strait and on western Seward Peninsula (Barker et al., 1994; D.M. Hopkins, pers. comm. 1997).

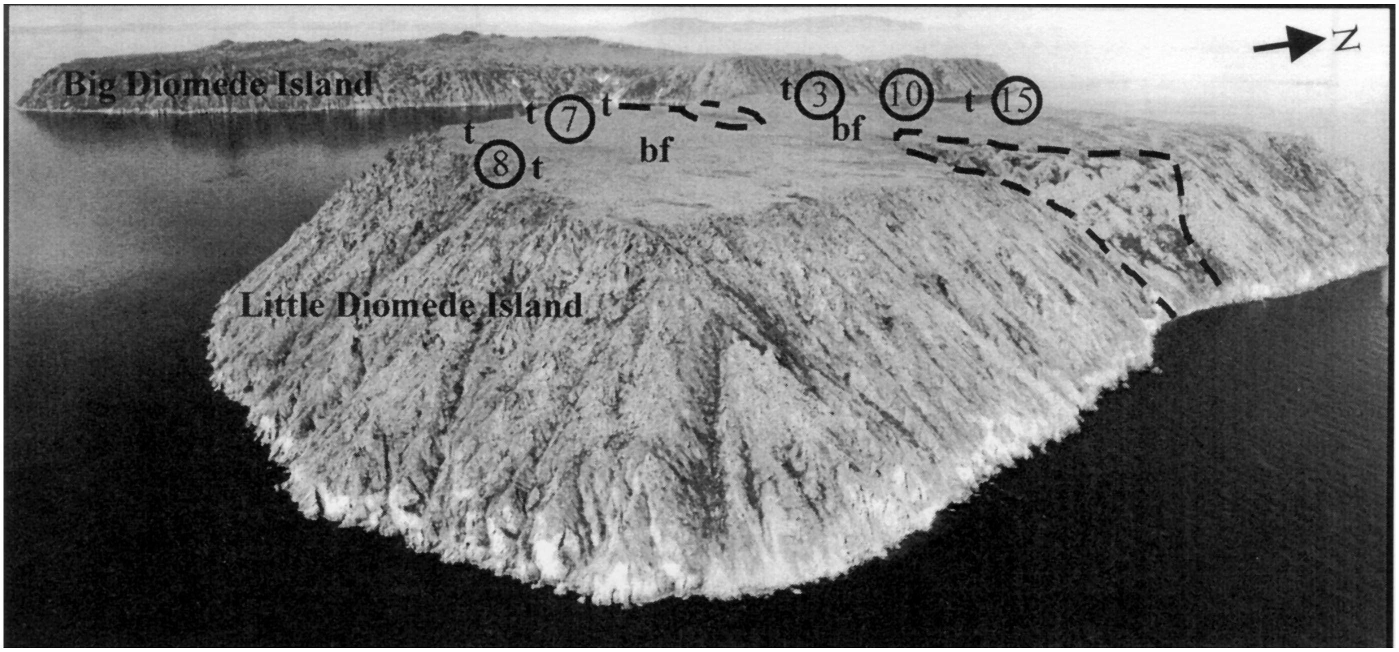

FIG. 2. Looking west across the Diomede Islands. t = areas where tors are found, bf = blockfield. Numbers in circles are ^{10}Be sample numbers preceded by the prefix 97LD. Dashed lines outline the two main valleys. On Big Diomede Island, several tors are visible as angular projections along the upper profile. Note the absence of shorelines or marine sediment on the cliffsides of the island. Air photograph provided by AeroMap U.S.

GEOLOGY OF THE BERING SEA AND BERING STRAIT REGION

The northern Bering Sea is a shallow, epicontinental sea (45% of the area is less than 200 m deep; Nelson et al., 1974; Sharma, 1974; Nelson, 1982). It has been separated from the Pacific Ocean by the Aleutian Islands since the late Eocene, when the Aleutian arc-trench subduction zone formed (Nelson et al., 1974; Worall, 1991). Subsequently, right lateral strike slip faults evolved parallel to the Beringian Active Margin; today these faults extend from the Sea of Okhotsk to southern Alaska (Worall, 1991). During a southward shift in the subduction zone, the Kula plate, of which the Bering Sea is a remnant, became incorporated into the North American-Siberian plate (Nelson et al., 1974). Movement of strike slip faults to the north and south of the Bering Strait determined its physiography, and played a crucial role in the intercontinental land connections between modern Asia and North America (Nelson et al., 1974). The Bering Strait Fault Zone, 38 km south of Little Diomede Island, trends west-east from the Bering Strait to Grantley Harbor (Fig. 1; Plafker et al., 1993).

In a study of marine sediments, Grim and McManus (1970) interpreted the surface sediments west of 169˚ W as glacial and marine in origin, whereas they interpreted the surface sediments east of 169˚ W to be underlain by river deposits. Glacial deposits have also been interpreted in the west part of the Chirikov Basin by Hess (1985). Grim and McManus (1970) also identified a ridge 75 km offshore, roughly paralleling Chukotka for 130 km, which is buried 30–40 m below the seafloor (Fig. 1). Boulders recovered in dredge hauls 9 km north of St. Lawrence Island suggest that the feature is a glacial moraine. This moraine is interpreted as Marine Isotope Stage (MIS) 8 in age (Krestaage, Russian terminology, D.M. Hopkins, pers. comm. 1997, Fig. 3). Toward its northern end, the moraine is defined by a belt of widely spaced chaotic seismic reflectors (possibly erratics), also interpreted to be glacial in origin (D.M. Hopkins, pers. comm. 1997). Although the inferred moraine ridge extends 150 km to the NNE from St. Lawrence Island, it terminates 50 – 100 km south of Little Diomede Island.

Tectonic activity in the Bering Strait since the late Tertiary/early Quaternary is indicated by deformed scarps, raised shorelines, and marine terraces in the York Mountains (Fig. 1) and offshore of Seward Peninsula (Grim and McManus, 1970; Kaufman and Hopkins, 1986). Differential uplift rates are estimated to be 5 m/km along the Port Clarence Fault Zone (Fig. 1; Sainsbury, 1967a).

The York Terrace

Numerous high sea level events ranging in age from late Pliocene to recent are recognized in western Alaska (Fig. 3; Hopkins, 1967; Kaufman, 1992; Kaufman and Brigham-Grette, 1993). The ages of the events are determined by stratigraphic relationships, distinctive molluscan faunas, position in geomagnetic polarity-reversal sequences, radiometric dating, and amino acid geochronology.

The York terrace is a 2.6 million-year-old marine surface, which represents a sea level stand 3060 m higher than the present level in the Bering Strait. The terrace is a prominent, continuous surface that can be traced for 30 km

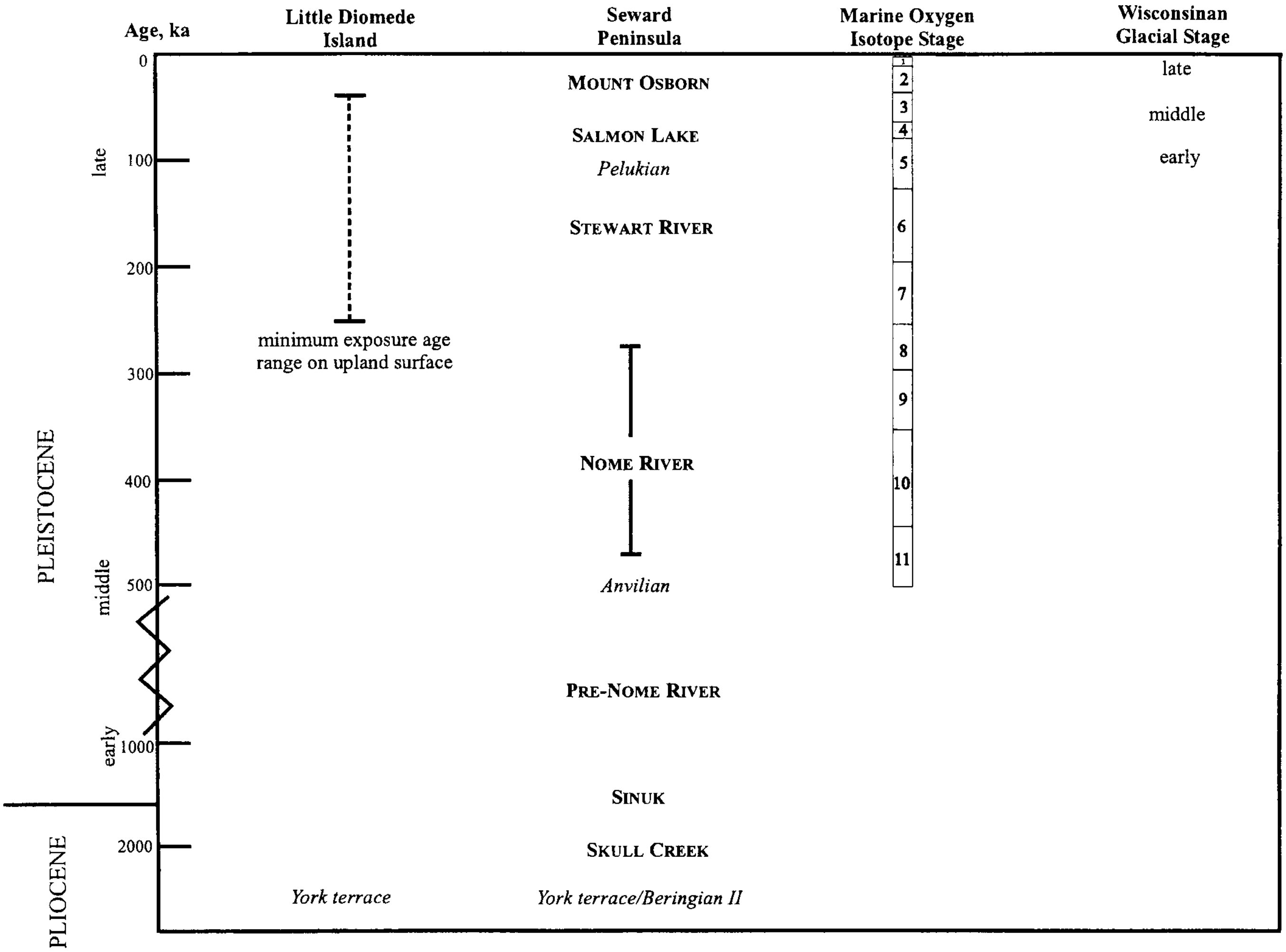

FIG. 3. Correlation of glacial and marine events on Little Diomede Island and Seward Peninsula. Glaciations in small capitals and sea level transgressions in italics. Note break in age scale.

along the southern front of the York Mountains (Sainsbury, 1967a; Brigham-Grette and Hopkins, 1994). It was originally dated at 2.6 Ma by mollusk faunas in marine conglomerate (Sainsbury, 1967a), and recent studies have confirmed the age on the basis of paleomagnetic data and the presence of the extinct mollusk *Fortipecten hallae*, a last appearance datum for this species in the region (Brigham-Grette and Hopkins, 1994). Amino acid analysis on *Mya* and *Hiatella* found in association with *Fortipecten hallae* from the California River valley and Nome support a correlation between the York terrace and the Beringian II marine transgression described in western Alaska (Kaufman and Brigham-Grette, 1993; Brigham-Grette and Hopkins, 1994). Subsequent tectonic activity during the mid Pleistocene uplifted the 30–60 m shoreline an additional 155–185 m along the south front of the York Mountains. The resulting altitude of the York terrace is 156 m on Cape Mountain and up to 225 m west of Lost River (Fig. 1; Sainsbury, 1967b). The extent of the uplift that deformed this shoreline in the Bering Strait was unknown before the present study, although it is critical to the Pliocene/early Pleistocene paleogeography of the Bering Strait. If the York terrace is projected 43 km west of Cape Mountain to Little Diomede Island using Sainsbury's (1967a) 5 m/km uplift rate, the elevation of the York terrace would be 371 m.

Regional Glacial History

In the absence of detailed studies on Little Diomede Island, the glacial history is discussed in the context of regional studies.

St. Lawrence Island: Holmes and Thor (1982) proposed an eastern limit for Chukotka Ice of unknown age extending onto St. Lawrence and King Islands (Fig. 1). Supportive evidence includes glaciotectonic folds and faults in sediment on the northwest corner of St. Lawrence Island and amino acid geochronology on mollusks incorporated into glacial deposits on the island (Hopkins et al., 1972; Benson, 1993, 1994). Recent studies have shown that ice advanced twice onto the island. The first advance occurred just after the mid Pleistocene (MIS 11) Anvilian marine transgression (Ivanov, 1986; Brigham-Grette et al., 1992). During MIS 5, small cirque glaciers formed at

155 m on St. Lawrence Island, after which, during late MIS 5 or MIS 4, ice from Chukotka is proposed to have advanced eastward onto the island (Heiser, 1997; Brigham-Grette et al., 2001). Projected snowlines for the late Wisconsinan (MIS 2) on St. Lawrence Island are 150 m in altitude, rising to 300 – 450 m in the Bering Strait (Heiser, 1997).

St. George and Aleutian Islands: Frost-shattered roches moutonnées, faceted and striated boulders, stratified drift, striae, and till indicate that a small ice cap and 2 – 4 cirque glaciers existed on St. George Island (Hopkins and Einarsson, 1966). Although there is no absolute age control for this glaciation, relative weathering criteria suggest it is correlative with the Nome River glaciation on Seward Peninsula, considered to be mid Pleistocene in age (Hopkins and Einarsson, 1966; Kaufman and Hopkins, 1986). Mid Pleistocene tills are also recognized on some of the Aleutian Islands (Thorson and Hamilton, 1986).

Seward Peninsula: Seven main glacial advances are recognized on Seward Peninsula (Figs. 1 and 3; Sainsbury, 1967b; Kaufman and Hopkins, 1986). The three earliest glaciations are discussed below. The pre-Nome River Drift (early to mid Pleistocene) is delineated by isolated erratics and extensive drift blankets on which little primary glacial relief is preserved. Evidence for the Sinuk glaciation (pre-Pleistocene to early Pleistocene) includes smooth moraine ridges south of the York Mountains at the mouths of the Kanauguk, King, and Lost Rivers, as well as till on the peninsula extending into northeast Lopp Lagoon (Fig. 1). The presence of tors on bedrock slopes on western Seward Peninsula originally suggested that the upland surfaces were glaciated only during the Sinuk interval; however, it has been demonstrated that tors can be preserved under cold-based as well as late Wisconsinan ice sheets (Sugden and Watts, 1977). The Sinuk glaciation moraine ridges extend 7 km offshore at Nome. The Skull Creek glaciation (pre-Pleistocene) is recognized by scattered erratics on low bedrock hilltops near Skull Creek and in the valley of the California River (Fig. 1), as well as by ice-rounded topography on hills between the western end of the York Mountains and Cape Mountain (Sainsbury, 1967b). The distribution of limestone drift at the northwestern end of the York Mountains suggests that an early Wisconsinan ice sheet occupied lowlands of the southernmost Chukchi Sea and Bering Strait area (Sainsbury, 1967a).

Bering Strait: A large portion of the research on past glaciation in the Bering Strait has been theoretical (Grosswald, 1988, 1998; Hughes and Hughes, 1994; Grosswald and Hughes, 1995). Grosswald (1998) proposed that sometime during the Pleistocene a marine-based ice sheet was centered over the East Siberian Sea, extending southward into the Bering Strait. According to this model, Little Diomede Island would have been covered by 1000 – 1500 m of ice. Sher (1995) and investigations by others using SAR imagery (Heiser, 1997), field-based mapping and numerical dating techniques (Gualtieri et al., 2000), and amino acid analyses on fossil mollusks from glaciomarine deposits (Brigham-Grette et al., 2001) indicate that this model is unsubstantiated.

METHODOLOGY

In July 1997, we investigated the perimeter of Little Diomede Island by boat and traversed the island's upland surface on foot. Five samples of granite from the upland surface were dated using ^{10}Be cosmogenic isotope analysis, a surface exposure dating technique (Fig. 2, Table 1). The premise of cosmogenic isotope surface exposure dating is that buildup and accumulation of cosmogenic isotopes begins once a rock is exposed at or near the surface. Using Accelerator Mass Spectrometry (AMS), the isotopic ratio of the cosmogenically produced isotope (in this case ^{10}Be) to the stable isotope (in this case ^{9}Be) is determined. The ratio is then multiplied by the concentration of the stable isotope in the sample, which is known from the mass and concentration of the Be spike added to the sample. Knowing the half-life of ^{10}Be (1.5 Ma), one can determine the length of time that the rock has been exposed at the surface (i.e., exposure age). For details regarding the use of cosmogenic isotopes as a surface exposure dating technique, see papers by Davis and Schaeffer (1955), Phillips et al. (1986), Lal (1991), Zreda et al. (1991), Nishiizumi et al. (1993), Kurz and Brook (1994), Zreda and Phillips (1994), Bierman et al. (1995), Clark et al. (1995), Gosse et al. (1996), and Bard (1997).

Five dated samples were taken by means of a hammer and chisel from the edge of bedrock outcrops. Sample locations are shown in Figure 2. Sample 97LD3, a slab 4 cm thick, was taken from the top of a granite outcrop (3 m long, 2 m wide, 15 m high) dipping 12˚ on the northwest side of the island. Sample 97LD7, a slab 7 cm thick, was taken from the uppermost surface of a tor dipping 32˚ on the southwest side of the island. The tor is 10 m long, 4 m wide, and 7 m high. Sample 97LD8, a slab 5 cm thick, was taken from the horizontal uppermost surface of an outcrop at the southern end of the island. Sample 97LD10, a slab 5 cm thick, was taken from an overhanging surface dipping 30˚ on an outcrop at the northern end of the island. Sample 97LD15, a slab 4.5 cm thick, was taken from a horizontal tor surface 3 m below the uppermost surface of the northern end of the island (Fig. 5).

Rocks were first cleaned using a wire brush to remove lichens. Samples were crushed, and the 500 – 250 mm size fraction was sieved off and then subjected to chemical pretreatment. Samples were cleaned with water and passed through a hand magnet to remove iron and other magnetic particles acquired during the crushing process. The final step in the physical sample preparation was to isolate the quartz. Quartz is used as the target mineral because it has the simplest target chemistry; it is also ubiquitous and chemically resistant. In addition, the use of quartz allows rigorous acid leaching to remove surface contamination

TABLE 1. Beryllium-10 exposure ages. AMS measurements were made on quartz from granite. Exposure ages are given for erosion rates of 0 and 10 mm/ka. Ages were calculated using PRIME Lab's RICH computer program.

Sample		NR/S[1]	no erosion RICH Age[2]	10 mm/ka RICH Age[2]
97LD3	blockfield	69 ± 9	41 ± 6	66 ± 17
97LD7	tor	155 ± 29	141.8 ± 28	254[3]
97LD8	blockfield	102 ± 17	80 ± 14	143[3]
97LD10	blockfield	37 ± 11	28 ± 8	36 ± 15
97LD15	tor	69 ± 11	47 ± 8	84 ± 31

[1] NR/S = The normalized radionuclide/stable nuclide ratio and reported absolute error. Units are in E-15. Errors on NR/S ratios and ages are limited by precision of AMS measurements at PRIME Lab. Production rate was corrected for latitude of 65˚N and an altitude of 363 m.

[2] Ages are in thousands of years. No shielding corrections were required for any samples.

[3] Extrapolation using an increase in "no erosion age" of 79%.

and atmospherically produced ^{10}Be. The quartz and other felsic minerals were separated using a Carpco™ Dry High-Intensity Induced-Roll Magnetic Separator.

After separation, the samples were etched with phosphoric acid two or three times, or until all other minerals (mostly feldspars) had been removed and only quartz remained. The "clean" quartz was then sent to Chemistry Operations at PRIME Lab (Purdue University) for the remainder of the chemical preparation. Complete chemical preparation procedures can be obtained from Purdue University's PRIME Lab Chemistry Operations Worksheets AW0010 and AW0011 (PRIME Lab, 1995). The chemical sample preparation involves leaching the quartz with hydrofluoric and nitric acid. AMS target preparation begins with adding a Be spike to the sample by producing a Be carrier. This step is followed by evaporation of the quartz, cation separation, cation purification, oxidation, and packing.

SURFICIAL GEOLOGY OF LITTLE DIOMEDE ISLAND

The modern upland surface of Little Diomede Island is a 350–363 m high frost-shattered surface sloping to the northeast with tors along its west and south sides (Fig. 2). The upland may be classified as a cryoplanation surface: a land surface reduced to low relief by processes associated with intensive frost action, supplemented by the actions of running water, moving ice, and other agents (Bates and Jackson, 1983; Allaby and Allaby, 1991). Reger and Péwé (1976) classify cryoplanation surfaces as bedrock steps or terraces on ridge crests and hilltops, chiefly occurring in nonglaciated areas near the general altitude of the snowline. Cryoplanation terraces are widespread on the uplands in nonglaciated portions of central Alaska, including the western tip of Seward Peninsula (Reger and Péwé, 1976). Reger and Péwé (1976) suggest formation of the surfaces is largely due to scarp retreat as the result of nivation.

Drainage Channels

The upper surface of the island supports one main valley on its east side (Fig. 2) and a smaller valley on its west side (Fig. 2). The remaining steep sides of the island are notched by small channels that drain small areas of the upland surface (Fig. 2). The larger eastern valley is approximately 75 m wide and can be traced from the eastern rim of the island to sea level. Drainage through the channel feeding the valley is ephemeral.

Visible from oblique air photographs are two former drainage channels cut into bedrock and oriented east-west. Solifluction lobes, consisting of weathered material, define a radial drainage pattern also visible from air photographs.

Blockfield

The blockfield is a transitional area between the east-side valley and the outer rim of tors and slabs of frost-shattered granite (Fig. 2). The blockfield is composed of coarse, angular granite blocks, up to 4 m in diameter. Some of the boulders are perched on smaller rocks (Fig. 4). It appears that the boulders in low spots have been frost-heaved to the surface, while most others close to the outer rim of the island have been frost-shattered or cracked in situ. The surface is similar in morphology to the "stone runs" and up-ended slabs of quartzite in the Falkland Islands as described by Clapperton (1993).

Tors

The morphology of some tors on the north and west outer rims of the upland surface resemble roches moutonnées (Fig. 5); however, striations or polished surfaces, if present in the past, are not preserved in the frost-shattered granite. The lee sides of these tors face west, implying that they were the product of glacial erosion from ice flowing from the east; however, there is no further evidence to support ice from the east (mainland Alaska) flowing onto Little Diomede Island. Alternatively, the form of the tors may be related to jointing patterns in the local bedrock.

Large and Small-Scale Weathering

Observations regarding the extent of differential weathering are especially important for interpreting cosmogenic isotope ages. Evidence for large-scale weathering includes blocks, some as large as 5 m in diameter, that have been heaved apart or broken at right angles to the outcrop. On the north side of the island, alongside the footpath to the upland surface, is a granite outcrop that displays at least 60 cm of onionskin exfoliation. In areas on the top of the island, it is also common to see sheets of granite 5 cm thick

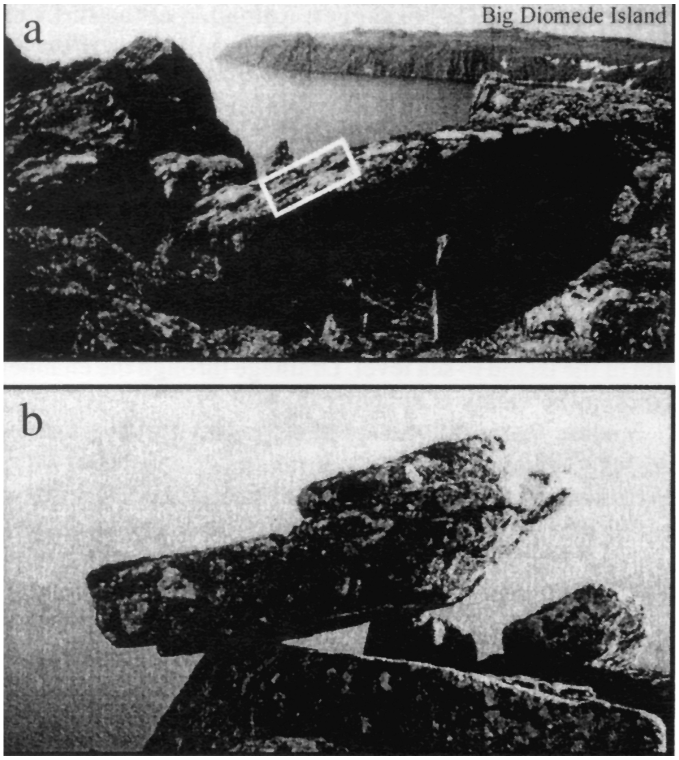

FIG. 4. a) Perched boulder on north side of the island (hammer in white rectangle indicates scale); b) Perched boulder on north side of the upper plateau ~350 m. Boulder is approximately 3 m in length.

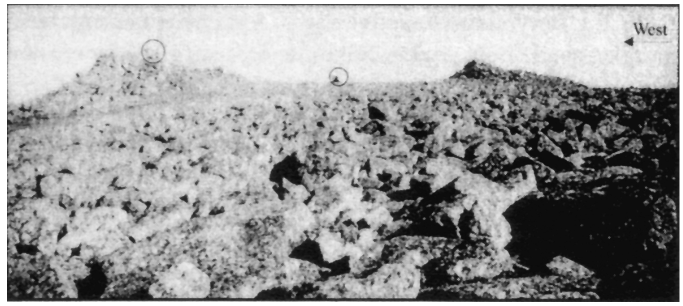

FIG. 5. Looking north at two roche moutonnée–shaped granite tors on the north rim of the island. The ^{10}Be cosmogenic isotope age on the westernmost tor assuming an erosion rate of 10 mm/ka is 84 ± 31 ka. There are two satellite dishes for scale (circled).

being broken apart from outcrops. Evidence for small-scale weathering includes exposed feldspar crystals and quartz veins standing up to 10 mm above the surrounding matrix. Granular disintegration in granite, producing hollows 5 cm deep, is also observed on flat-topped blocks on the north and west rims. Moss, which is 10 cm thick on some rocks, is present to trap water and facilitate mechanical erosion.

DISCUSSION OF THE AGE AND ORIGIN OF THE UPLAND SURFACE

The exposure age of the upland surface on Little Diomede Island, determined by ^{10}Be cosmogenic isotope dating, ranges from 36 to 254 ka (Table 1). Our hypothesis is that the island's upland surface evolved in two stages: 1) a late Pliocene marine planation followed by tectonic uplift and 2) a period of periglacial activity.

Cosmogenic Isotope Dates and Reasons for Age Distribution

It is difficult to assess the exact age of the upland surface because of variability within the cosmogenic isotope ages. Disparate ages result from a combination of analytical errors (associated with the chemical sample preparation and the accelerator mass spectrometry measurements at PRIME Lab) and geologic factors (such as differential weathering and the dip of the sampled surface). However, because of weathering and possible underestimation of the production rate of ^{10}Be, the dates are interpreted to be minimum ages. Since both large- and small-scale weathering features are present in the granite, it is difficult to assess how much rock has been removed by weathering and erosion since the rock surfaces were first exposed. Therefore, the ages are interpreted to be minimum exposure age estimates.

A range of ages is given using a "no erosion" model and a "10 mm/ka" erosion model (Table 1). An erosion rate of 10 mm/ka was chosen on the basis of the maximum relief of quartz veins (10 mm) on the sampled surfaces, which provides evidence that at least 10 mm of erosion has taken place. It was not possible to calculate a 10 mm/ka erosion rate for older samples (more than 80 ka old) because of limitations in RICH, PRIME Lab's age calculation program. However, if the same percentage increase (79%) as that of the next oldest sample (97LD15, 47 ± 8 ka) is used to calculate a "10 mm/ka" erosion rate for samples 97LD8 (older than 80 ka) and 97LD7 (older than 142 ka), their ages become 143 and 254 ka, respectively (Table 1). Using this extrapolation, the "10 mm/ka" erosion model ages range from 36 to 254 ka. The "no erosion" ages are unrealistic, since signs of weathering are ubiquitous on the rocks; therefore, the range of ages of the five samples assuming 10 mm/ka erosion (36 – 254 ka) is used for interpretation.

Stage I (Late Pliocene/Early Pleistocene) – Marine

As the elevation of the upland surface of Little Diomede Island is 363 m, the hypothesis that the upland surface of Little Diomede Island is an extension of the York terrace is supported by uplift rates and regional-scale geomorphology. The planar upland surface of Little Diomede Island is probably best interpreted as a former marine wave-cut platform. If the York terrace was uplifted 2.6 million years ago and has been eroding at the rate of 10 mm/ka, then it has undergone a total of 26 m of erosion. As was noted earlier, the projected elevation of the York terrace on Little Diomede Island is 371 m; however, after 26 m of erosion this would be reduced to 345 m, which is close to the present elevation (350–363 m) of the upper surface of the island. Using Sainsbury's (1967a) uplift rates, we conclude that this surface may be a remnant of the 2.6 million-year-old York terrace in the Bering Strait. However, it is also possible that there are numerous other terraces between the York terrace and the Lost River terrace and that the Little Diomede upland represents one of these intermediate high sea level stands (T. Hudson, pers. comm. 1998). The correlation is based on geomorphology, projected uplift rates, and assumed erosion rates. Note that the rounded pebbles or beach shingle common on other raised marine surfaces in the Arctic are not found on the upland surface of Little Diomede Island.

Stage II (Late Pleistocene to Present) – Periglacial

Subsequent to marine planation, the Little Diomede Island plateau has likely been dominated by periglacial weathering processes for at least 36 ka (since MIS 3) and possibly 254 ka (since MIS 7/8). Evidence for subaerial erosion throughout the Pleistocene on the exposed Bering shelf has also been recognized (Grim and McManus, 1970). Evidence for the periglacial environment on Little Diomede Island includes the cryoplanation surface, the blockfield, and the tors. The length of time required for blockfields to form varies with rock type and climate, but within the Bering Strait, this process may be accelerated by high amounts of precipitation, fog, frost action, and salt weathering (Embleton and King, 1975).

The Possibility of Glaciation?

Although there is no direct evidence for glaciation of Little Diomede Island, it is possible that perennial snowbanks or even a small plateau ice cap existed on the island during the early Pleistocene. In support of glaciation of the island are the valleys on the east and west sides, which may be remnants of former drainage, as well as the perched boulders (Fig. 4), which could be interpreted as glacially deposited. The early Pleistocene snowline dropped to 150 m on St. George Island, south of Little Diomede Island, and it was probably lower than 200–250 m on Seward Peninsula (Hopkins and Einarsson, 1966; Kaufman and Hopkins, 1986). On the basis of estimated snowline elevation alone, Little Diomede Island fits into the regional pattern of westward sloping paleosnowline altitudes reconstructed for the Bering Strait and the York Mountains during the late Pleistocene. Offshore seismic evidence and field observations on Little Diomede Island (i.e., no erratics found on the island, west-facing lee side tors) indicate it is unlikely that ice from Chukotka actually reached the island.

CONCLUSIONS

On the basis of geomorphology and the projection of uplift rates, we hypothesize that the flattened top of the upland surface of Little Diomede Island is a former marine terrace correlative with the 2.6 million-year-old York terrace on Seward Peninsula. After marine planation, Little Diomede Island developed a cryoplanation surface because of intense periglacial weathering at least since MIS 3 and possibly since MIS 7/8 (Fig. 3). More cosmogenic isotope dating on tors and the blockfield from the upland surface of Little Diomede Island and an investigation of the surficial geology of Big Diomede Island are needed to further constrain the age of the two stages. The implications for this research are the following:

1. If the recognition and projection of the York terrace in the Bering Strait is correct, this evidence helps to constrain early Pleistocene uplift rates and regional warping in the Bering Sea.

2. Evidence indicates that the upland surface of Little Diomede Island has remained a periglacial environment for at least 36 ka; therefore, it was not covered by ice due to extensive glaciation. Thus this work adds to the continuum of research substantiating restricted ice cover and modest sea level fluctuations in the Bering Strait region since the late Pleistocene.

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