Alpine glacial geology of the Tablelands, Gros Morne National Park, Newfoundland

Gerald Osborn, Ian Spooner, John Gosse, and Doug Clark

Abstract: Controversy persists in western Newfoundland regarding Pleistocene, particularly Late Wisconsinan, glacial ice volumes. Independently, a set of alpine glacial deposits on the flanks of the Tablelands in Gros Morne National Park has attracted much attention but little scrutiny. In this study, cosmogenic nuclide dating of the alpine deposits places some limits on post-late glacial maximum (LGM) ice dynamics in the vicinity of the Tablelands, a plateau bounded on the northeast by Trout River Gulch. Small valleys incised into the flanks of the Tablelands are floored with a diamict that contains both till and ice-contact deposits. Rock glaciers rest on the diamict, and rock glacierization also has affected talus lining the south wall of Trout River Gulch. A small moraine rests in the Devil's Punchbowl cirque. The cirque moraine, lobate deposits below the cirque moraine, rock glaciers, and a colluvial veneer overlying the till in the small valleys have cosmogenic ^{36}Cl ages as old as either ca. 20 or 15 ka, depending on what erosion rate is assumed, indicating that these bodies are Late Wisconsinan in age but post-date the local LGM. Trout River Gulch was deglaciated early and perhaps did not contain active ice even at the LGM, but previous work shows that ice was streaming seaward both north of Trout River Gulch and south of the Tablelands even as the gulch lay relatively ice free.

Resumé: Il subsiste des controverses dans l'Ouest de Terre-Neuve quant au volume des glaces au Pléistocène, surtout au Wisconsinien tardif. Indépendamment, un ensemble de dépôts de glaciers alpins sur les flancs des Tablelands dans le parc national Gros Morne a attiré beaucoup d'attention mais peu d'analyse. Dans cette étude, la datation de dépôts alpins ayant des nucléides cosmogéniques établit certaines contraintes sur la dynamique des glaces après le dernier maximum glaciaire aux environs des Tablelands, un plateau limité au nord-est par le ravin de la rivière Trout. Les fonds de petites vallées coupant les flancs des Tablelands sont tapissés d'un diamicton qui contient du till et des dépôts de contact glaciaire. Des glaciers rocheux reposent sur le diamicton et l'englaciation des roches a aussi affecté les talus qui tapissent le mur sud du ravin de la rivière Trout. Une petite moraine réside dans le cirque Devil's Punchbowl. La moraine de cirque, des dépôts en forme de lobes sous la moraine de cirque, des glaciers de roches et un placage de sédiments colluviaux par-dessus le till dans les petites vallées ont des âges cosmogéniques ^{36}Cl aussi anciens que 20 ou 15 ka, selon le taux d'érosion assumé, indiquant que ces amas datent du Wisconsinien tardif mais sont plus tardifs que le dernier maximum glaciaire local. Le ravin de la rivière Trout a été déglacé tôt et il ne contenait peut-être pas de glace active même au dernier maximum glaciaire; cependant, des études antérieures ont démontré que la glace s'écoulait vers la mer à la fois au nord du ravin de la rivière Trout et au sud des Tablelands alors même que le ravin demeurait relativement libre de glace.

[Traduit par la Rédaction]

Introduction

There are tremendous differences in paleoglacier ice volume estimates for regions of Canada that are easily accessed and already mapped (Marshall et al. 2002). Contributing to the disparities are (*i*) the lack of adequate field data in many areas with which to constrain ice sheet models; (*ii*) the limited nature of glacial geochronology; (*iii*) our growing sense of low-erosive ice and our uncertainty in the conditions causing it (Staiger et al. 2005); and (*iv*) the growing weight of evidence for biological refugia to explain speciation and disjunct species distributions (Rogers et al. 1991). One of these regions is western Newfoundland, where glacial ice cover is particularly difficult to decipher because of its proximity to the ocean, high coastal relief, and peripheral position along the eastern margin of the Laurentide Ice Sheet (LIS). Despite nearly a century of field observations, surficial mapping of part of the region at a scale of 1 : 50 000, low-resolution,

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offshore seismic and multibeam bathymetry, marine and terrestrial radiocarbon chronology, and recent measurements of terrestrial cosmogenic nuclide (TCN) exposure ages, the controversy has become even stronger. The debate hinges largely on the lack of means by which to measure ice thickness in key areas.

Exposures of ice-marginal features on land provide an opportunity to delimit ice volume in southern Gros Morne National Park in western Newfoundland (Fig. 1). We have combined sediment stratigraphy, geomorphology, and TCN exposure ages of boulders on small cirque moraines, along with previous observations, to place some constraints on ice cover in the vicinity of a prominent plateau known as the Tablelands. We address as part of the argument the age and origin of small-scale alpine glacial features of the Tablelands, which have attracted much attention but never before have been studied in detail.

Physical setting

Gros Morne National Park is located in the Long Range Mountains (Fig. 1), a long (!), narrow, flat-topped upland divided by fluvial and glacial erosion into alpine plateaus and rounded summits. The plateaus rise 400–800 m above sea level. Deep fjords cut the Long Range Mountains; some of them are cut off from the sea by moraines and glaciomarine deltaic deposits and (or) bedrock ridges and form long, narrow lakes locally known as ponds. The best known of these in the study area is Trout River Pond (Fig. 1). Coastal lowlands border the Gulf of St. Lawrence. A generally cool, wet climate generates boreal forest, studded with occasional peat bogs, in the lowlands; the uplands are mantled with tundra and swamps. Even tundra vegetation is scarce where the exposed bedrock is peridotite.

Rocks in the Gros Morne region record Proterozoic and early Paleozoic collisional tectonics and deposition in the Iapetus Ocean. The bulk of the Long Range Mountains in the park is underlain by Proterozoic granite and gneiss (Berger et al. 1992). Adjacent to the coast, a 5–10 km wide strip of land is underlain by continental-margin clastic and carbonate sediments, along with collisional mélange. In the southern part of the park, the subject of this paper, ophiolitic peridotite and gabbro are prominent.

The Tablelands (generally expressed in the plural) refer to a steep-flanked, rolling plateau rising to 711 m, south of the Bonne Bay fjord system (Figs. 1, 2). Brookes (1993) calls the plateau “Table Mountain.” The edges of the plateau are steepest and best defined in the north, where the margin is defined by a once glaciated valley known as Trout River Gulch, and in the southwest, where Trout River Pond fills the bottom of a fjord valley that is designated Trout River Pond Valley (Fig. 2). Coastal lowlands border the plateau on the northwest, and hills of intermediate elevation on the east-southeast. The plateau is underlain by ophiolitic rocks of the Humber Arm Allochthon (Williams and Cawood 1989). Gabbro is exposed in the northwestern third of the upland, but elsewhere peridotite is at the surface. Plant growth on the latter rock is very limited by high Mg content, Ni toxicity, and nutrient scarcity (Roberts 1980). The Tablelands are characterized by barren, frost-shattered rock, brown-weathering where peridotite is exposed and somewhat more vegetated on gabbro surfaces. Soils are thin to non-existent, and evidence of active cryoturbation is common.

The summit plains of the Tablelands, along with similar surfaces elsewhere in the Long Range Mountains, generally are regarded as remnants of a planation surface(s) uplifted in Tertiary time (Brookes 1964, 1993; Grant 1989*a*; Berger et al. 1992), although low-temperature thermochronology suggests these surfaces may be Mesozoic in age (Hendricks et al. 1993).

Previous work

Reviews or summaries of the progress of Quaternary studies in Newfoundland are given by Tucker (1976), Grant (1977, 1989*a*), Brookes (1982), and Rogerson (1982). Submarine moraines off the west coast of Newfoundland most recently were addressed by Shaw (2003), and a large-scale view of deglaciation of Atlantic Canada is that of Shaw et al. (2006). Various studies cited in the latter paper place the late glacial maximum (LGM) in a range from ca. 21 000 to ca. 25 000 ^{14}C BP (ca. 25.2–30 cal ka),[2] but the Gulf of St. Lawrence ice margin (to which western Newfoundland ice drained) is shown as remaining unchanged through ca. 19 cal ka. Ice is shown as breaking up in the Gulf of St. Lawrence by ca. 16.8 cal ka, and the west coast of Newfoundland was exposed, at least in part, by ca. 14.8 cal ka.

Particularly germane to this study is the debate over Late Wisconsinan ice limits in western Newfoundland. Altitudinally separable weathering zones have been described in the Long Range Mountains (and in other coastal highlands in Newfoundland and Labrador); the most weathered zones are typically the highest, with lower zones being consecutively less weathered. According to one school of thought (e.g., Brookes 1977; Grant 1977), the weathering zones represent glacial extents, and the ages of the zones correspond to the timing of their last glacierization. The felsenmeer and tors on the high plateaus such as the Tablelands are taken to indicate that glacierization there pre-dated Late Wisconsinan time if indeed the plateaus ever were glaciated (e.g., Grant 1989*a*). The plateaus thus were Late Wisconsinan nunataks, and ice then occupied valleys and fjords. Meanwhile, TCN dating at sites in the Long Range Mountains (not including the Tablelands) suggests that Late Wisconsinan ice *did* cover the highlands (Gosse et al. 2006). These results are taken to mean that highland weathering zones represent different basal thermal regimes of glacier ice, rather than vertical ice limits, and felsenmeer are considered to have survived one or more glaciations (Marquette et al. 2004; Staiger et al. 2005; Gosse et al. 2006).

The Late Wisconsinan nunatak model was applied to the Tablelands in surficial mapping by Grant (1989*b*; more or less repeated by Berger et al. 1992). Brookes (1993) noted quartzite and granite erratics on the Tablelands (as well as peridotite on the gabbroic part of the surface) but regarded glacierization there as pre-dating the last two glaciations. He

[2] In this paper radiocarbon ages are reported as ^{14}C BP, calibrated radiocarbon ages as cal ka, and TCN exposure ages as ka.

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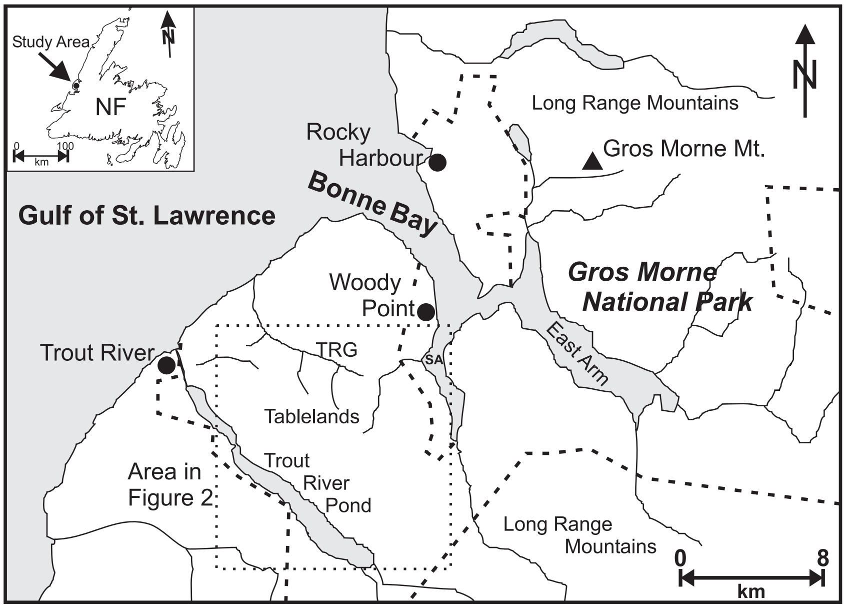
Fig. 1. The study site (dotted line, see Fig. 2) is located in western Newfoundland within Gros Morne National Park. The broken lines denote the park boundary. SA, South Arm of Bonne Bay; TRG, Trout River Gulch.



regarded Late Wisconsinan ice to have been restricted to flanking troughs.

The most noticeable alpine glacial forms in the area are in the road-accessible Trout River Gulch, on the north side of the Tablelands. They are mapped on surficial maps (Grant 1989*b*; Berger et al. 1992) and described in a short paper by Brookes (1993) and in a series of field trip guidebooks. There is no debate regarding the glacial erosion of the gulch, but the age of glacierization has been disputed, and no numerical dates were available prior to this study. Brookes assumed that valley glacier(s) in the gulch pre-dated Late Wisconsinan time because (*i*) a "lateral moraine" (discussed later in the paper) in the gulch contains quartzite and other erratics from 20 km or more east of Bonne Bay, deposited by ice that flowed westward across Bonne Bay; and (*ii*) Late Wisconsinan ice flowed northward through Bonne Bay (Fig. 1), so the west-directed gulch ice must have preceded that. These ideas were not elaborated on. Conversely, a Late Wisconsinan age for the last Gulch glacier is promoted by Grant (1987, 1989*b*), Proudfoot et al. (1988), Berger et al. (1992), Canadian Quaternary Association – Canadian Geomorphological Research Group (1995), and Batterson et al. (2001). The latter four are guidebook articles, and the later of these basically repeat material from the earlier. In addition to the lateral moraine, a cirque moraine in the Devil's Punchbowl cirque and a collection of rock glaciers (Figs. 2, 3) figure prominently in the guidebooks and maps listed previously; details are given in the appropriate sections later in the paper. Also addressed are some claims of glacial geomorphological features that we were unable to find.

Trout River Pond, the northwest segment of which is called Trout River Small Pond (Fig. 3), is blocked at its seaward end by an ice-marginal marine delta. *Hiatella* (sp.) shells from the deltaic sediments yielded a radiocarbon age of 12 400 ± 170 ^{14}C BP (GSC-4295), or 13.8 ± 0.2 cal ka (calibrated with CALIB 5.0.2), and a bowhead whale vertebra from the sediments was dated at 11 720 ± 205 ^{14}C BP (S-2149), or 13.2 ± 0.2 cal ka (Brookes 1974; Grant 1987). As these dates are associated with ice that was to the south of the Tablelands, they do not bear necessarily on the question of whether Late Wisconsinan ice occupied Trout River Gulch to the north of the Tablelands.

There has been little formal investigation of lake sediment or organic records in the Tablelands region. A sediment core from Long Pond (Fig. 2) yielded a date of 8950 ^{14}C BP (GSC-4659; McNeely and McCuaig 1991) from 413–421 cm depth, well above the base of the core at 752 cm. Anderson and McPherson (1994) indicate that the oldest basal lake-sediment date obtained from anywhere in southern or western Newfoundland is 13 400 ^{14}C BP (GSC-3559; Anderson 1983), from the south coast 300 km southeast of the Tablelands.

Geographic confusion

There are some nomenclatural inconsistencies in the previous literature that are addressed here for the sake of clarity (see Figs. 1–3).

The term Devil's Punchbowl does not appear on federal topographic maps; usage of it here is adopted from the guidebook of Grant (1987), later guidebooks, and Brookes (1993). Canadian Quaternary Association – Canadian Geomorphological Research Group (1995) and Batterson et al. (2001) mistakenly apply the term to an incised creek ("Berry Barren Brook" according to a sketch in Berger et al. 1992) west of the Punchbowl, but the original map (in Brookes 1993) from which the erroneous maps were reproduced correctly locates the Punchbowl.

Grant (1987), Canadian Quaternary Association – Canadian Geomorphological Research Group (1995), and Batterson et al. (2001) all state "Small cirque moraines … occur at the head of Winterhouse Gulch, and are especially clear in the next basin toward Trout River (Devil's Punchbowl)." But the next basin toward Trout River is not Devil's Punchbowl, rather it is a small valley 1.5 km east of the Punchbowl,

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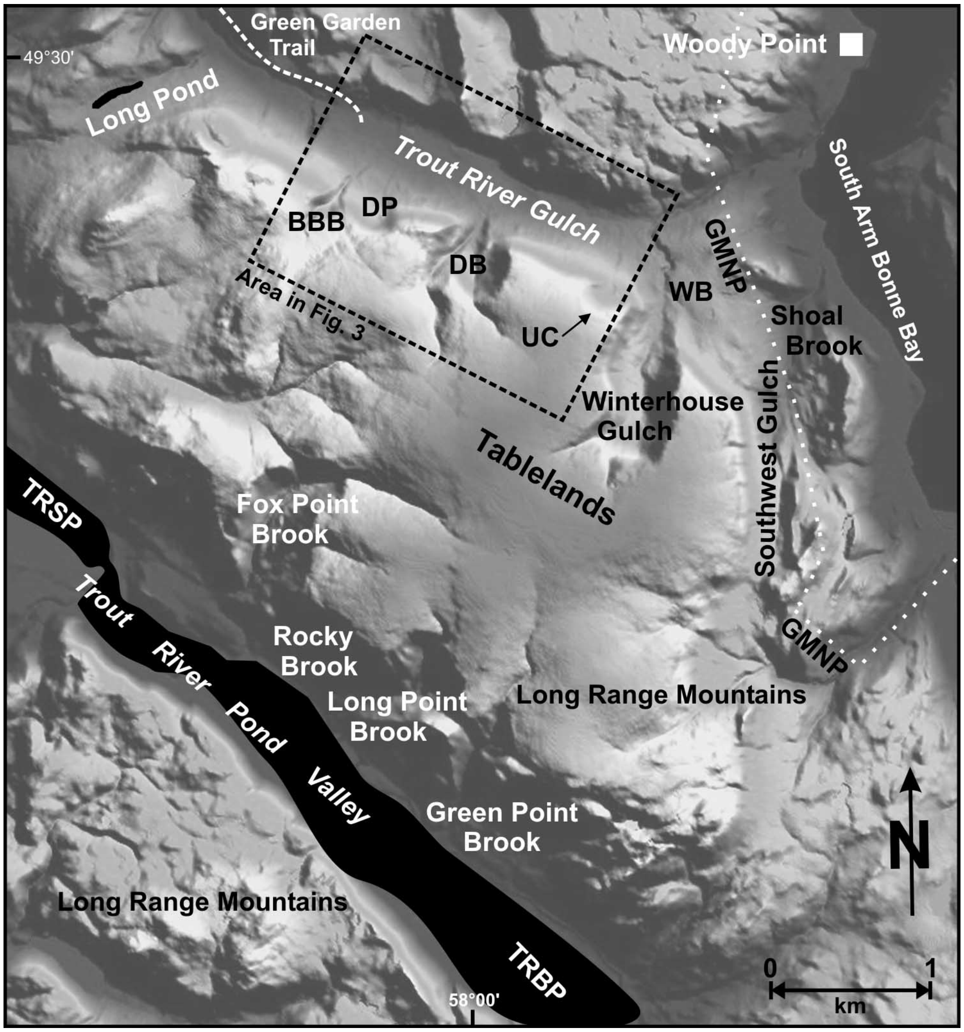
Fig. 2. Digital elevation model (10 m resolution) of the Tablelands in Gros Morne National Park (GMNP). BBB, Berry Barren Pond; DB, Dry Brook; DP, Devil's Punchbowl; TRBP, Trout River Big Pond; TRSP, Trout River Small Pond; UC, unnamed cirque; WB, Winterhouse Brook.



unlabeled on federal topographic maps. This small valley is labeled Dry Brook in a sketch in Berger et al. (1992) but is labeled Penman's Brook by Brookes (1993). A.R. Berger maintains (personal communication, 2003) that Dry Brook is the actual name, so it is used in this paper. To further complicate matters, Dry Brook is in fact wet.

Winterhouse Gulch is used by Grant (1987) and later guidebook writers to refer to the incised valley (Fig. 4), tributary to Trout River Gulch, occupied by the upper part of Winterhouse Brook. A.R. Berger (personal communication, 2003) maintains the correct term for the valley is "Winterhouse Brook." In this paper we continue to use "gulch" for the valley to avoid confusion between valley and the brook itself. Winterhouse Brook is also the name of a village 1 km south of Woody Point (Fig. 1).

Methodology

Fieldwork was conducted in the summers of 2002 and 2003. Mapping of glacial–periglacial deposits was done at a scale of 1 : 25 000 with the aid of airphotos.

Cosmogenic ^{36}Cl was used to determine the exposure age of boulders on the surfaces of the alpine deposits. TCNs such as ^{36}Cl are produced during interactions between cosmic rays and exposed minerals. The TCN concentration depends on the cosmic ray flux to the sampled surface (which is controlled by atmospheric shielding and geomagnetic latitude), the duration of exposure, and environmental factors such as erosion of the surface, exhumation of the boulder after deposition, boulder movement, and shielding due to snow cover. ^{36}Cl was selected because of the lithology of the boulders, which was largely peridotite cumulates of the Tablelands ophiolite. It was believed that the age (>400 Ma) and degree of serpentinization of the olivines precluded the use of cosmogenic ^{3}He. ^{36}Cl is produced from a number of interaction pathways, including spallation by fast nucleons with larger mass nuclei (mass number A > 36), thermal neutron capture (by ^{35}Cl), and muonic interactions. The ^{36}Cl can be measured in individual minerals or whole rock samples because production rates of the isotope have been calibrated for most of the target elements (although Fe and Ti are not yet well calibrated), and the thermal neutron cross section of

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Fig. 3. Location of sites discussed in the text and cosmogenic dating sites. The numbers refer to the last number for each sample location in Table 1. DP, Devil's Punchbowl; IDB, irregular debris bench along the edge of Trout River Gulch, interpreted by previous workers as a lateral moraine; RG, rock glacier. The shaded areas are uplands with elevations of greater than 500 m.

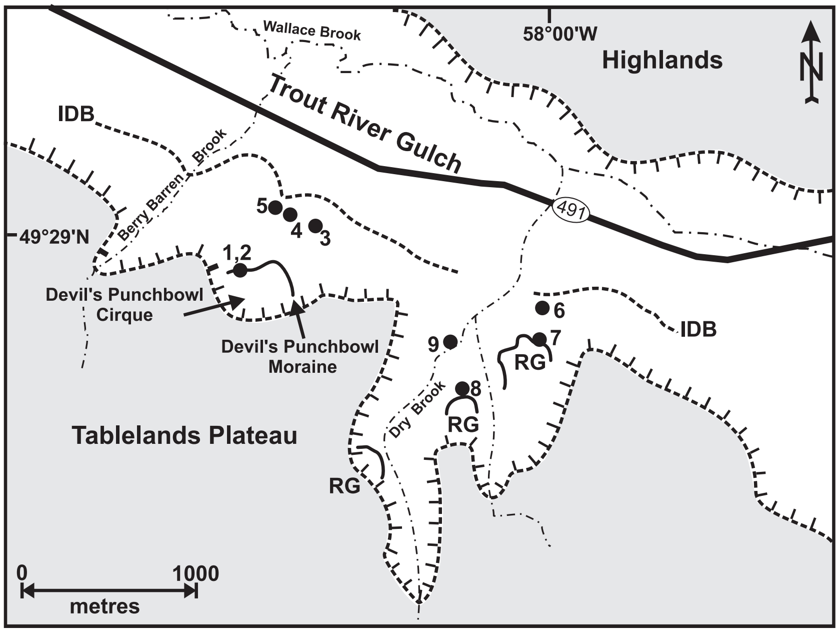

Fig. 4. Looking down the glaciated gorge of Winterhouse Brook, cut in peridotite. A thin layer of grey diamict, shown in Fig. 10, crops out intermittently along the floor of the valley.

^{35}Cl is reasonably well established (Phillips et al. 1996; Stone et al. 1996, 1998). For a review of the TCN technique and applications for dating glacial deposits, see Lal (1991) and Gosse and Phillips (2001).

We sought boulders, which have an exposure history that should date the time of initial exposure of the landform surface. Over the entire field area, only eight boulders were found to be reasonably suited for exposure dating (Table 1). None of the boulders had weathering rinds. We avoided boulders on slopes to preclude sampling a surface on a rolled boulder and excluded boulders near the bases of cliffs. Erosion of the boulder will change the TCN concentration of a rock surface, therefore it is important to minimize the possibility of sampling a boulder that has significantly eroded. Recognizing that not all boulders have the same erosion rate or style (e.g., some erode by grussification, spalling, or dissolution), we used the following approach to select boulders: (*i*) avoid boulders with large gnammas or rillen and other indications of substantial chemical weathering and dissolution; (*ii*) select boulders that have continuous sharp edges or flat upper surfaces (e.g., formed by a serpentinite vein); (*iii*) if possible, select boulders with surfaces that exhibit striae or other indicators of negligible erosion (no striae were found); and (*iv*) examine the material at the base of a boulder to find evidence of excessive spalling or grussification. Additional selection criteria included height (to minimize snow effects and neutron loss effects), simplicity of geometry (>1.5 m^2 flat-top boulders require no geometry corrections, otherwise adjustments would be required for sloped or peaked surfaces, or samples collected near the edge), and degree of weathering. Samples were never within 10 cm of an edge. Boulder heights averaged 81 cm, ranging from 45 to 170 cm, and were not optimum for exposure dating owing to the possibility of snow shielding in the past. However, the total uncertainty in each age we report reflects the uncertainty in erosion, snow cover, and other factors.

Samples averaging 1 kg in weight and 1–4 cm in thickness were collected from the tops of eight boulders on the Devil's Punchbowl cirque and Dry Brook deposits. Samples were prepared using standard isotope dilution methods for ^{36}Cl. Physical processing, water-content measurements, and data reduction and interpretation were completed at Dal-CNEF Laboratory at Dalhousie University, Halifax, Nova Scotia. TCN extraction was completed by T. Thomas at New Mexico Institute of Mining and Technology, Soccoro, New Mexico. Selected major, minor, and trace elements were measured by Chemex, and B and Gd were measured by the Canadian company Xral Laboratories, Mississauga, Ontario (Table 2). The ^{36}Cl/^{35}Cl ratio was measured by accelerator mass spectrometry at PRIME Lab, Purdue University, West Lafayette, Indiana.

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Table 1. Location and height for Trout River Gulch samples.

	Location				
Sample No.	Lat. N (°)	Long. W (°)	Elev. (m)	Height (cm)	Physical location
NF-02-PBC-201	49.47965	58.01975	330	90	West end of Devil's Punchbowl cirque moraine
NF-02-PBC-202	49.47965	58.01985	330	65	West end of Devil's Punchbowl cirque moraine
NF-02-PBC-203	49.48370	58.08373	230	70	Devil's Punchbowl lobate deposits
NF-02-PBC-204	49.48443	58.01757	210	170	Devil's Punchbowl lobate deposits
NF-02-PBC-205	49.48497	58.01918	211	45	Devil's Punchbowl lobate deposits
NF-02-PEN-206	49.47797	57.99363	222	80	On grey fill in front of Dry Brook rock glacier
NF-02-PEN-207	49.47495	57.99382	310	60	On Dry Brook rock glacier
NF-02-PEN-208	49.47138	57.99712	340	70	On till ridge upstream of Dry Brook rock glacier
NF-02-PEN-209	49.47507	57.99887	209	—	Shielded bedrock at Dry Brook confluence

Note: All samples are boulders except NF-02-PEN-209, which is a shielded >5 m deep bedrock sample.

Table 2. Chemical data for Trout River Gulch samples.

Sample No.	C (wt.%)	Na_2O (wt.%)	MgO (wt.%)	Al_2O_3 (wt.%)	SiO_2 (wt.%)	P_2O_5 (wt.%)	K_2O (wt.%)	CaO (wt.%)	TiO_2 (wt.%)	MnO (wt.%)	Fe_2O_3 (wt.%)	B (ppm)	Gd (ppm)
NF-02-PBC-201	7.60	0.03	39.63	1.11	42.04	0.01	0.01	0.79	0.01	0.11	8.72	nd	nd
NF-02-PBC-202	9.04	0.03	38.66	0.94	41.48	0.01	0.01	0.84	0.01	0.11	7.85	5.3	1.1
NF-02-PBC-203	11.01	0.03	37.60	0.56	40.96	0.01	0.01	0.45	0.01	0.11	8.33	nd	nd
NF-02-PBC-204	8.73	0.03	36.61	0.87	43.49	0.01	0.01	0.91	0.01	0.11	7.99	4.7	nd
NF-02-PBC-205	5.87	0.02	36.66	1.35	45.89	0.01	0.01	1.39	0.01	0.11	7.88	nd	nd
NF-02-PEN-206	10.11	0.02	39.09	0.74	39.81	0.01	0.01	0.71	0	0.10	8.92	nd	nd
NF-02-PEN-207	11.85	0	38.47	0.71	39.42	0.01	0.01	0.39	0	0.10	8.44	1.2	nd
NF-02-PEN-208	11.19	0.02	38.75	0.76	39.78	0.01	0.01	0.64	0.01	0.10	8.68	nd	nd
NF-02-PEN-209	10.94	0.02	39.79	0.66	39.36	0.01	0.01	0.30	0.03	0.10	7.96	4.6	nd

Note: Concentrations of U and Th in these ultramafic rocks were below the lower limit of detection by X-ray fluorescence (with the exception of ~1 ppm Th in sample NF-02-PBC-201). nd, below lower limit of detection for B and Gd by prompt γ emission spectrometry.

Exposure ages were calculated using production rates according to Phillips et al. (1996) using calculations presented in Chloe3 (Phillips and Plummer 1996). Uncertainties in the ages reflect the 1σ precision in the accelerator mass spectrometry (AMS) $^{36}Cl/^{35}Cl$ measurements. Adjustments for water content were found to be normal but significant (Table 3). Adjustments in the exposure age for the effects of boulder surface erosion are required owing to the sensitivity of ^{36}Cl exposure dating to erosion. Unlike other TCN systems that yield lower concentrations with higher erosion rates, the unique thermal neutron component of ^{36}Cl results in a nonlinear age function with erosion. ^{36}Cl concentrations measured in the surface of a boulder will increase under slow erosion rates (making unadjusted ages too old) but decrease under much faster rates.

No correction for vegetation was required because the ophiolite is toxic to most plants and has remained barren. Snow effects are believed to be small (<10%), based on (*i*) observations of snow cover of the moraines in the winter of 2003–2004, (*ii*) the low elevation of the sites, and (*iii*) the position of the boulders on ridge crests that probably would have remained wind blown for most of the 3 snow months. We cannot preclude the possibility that avalanches may have buried the boulders and caused significantly more shielding, although we have no data to support this possibility and most of the samples are located in positions that will minimize the influence of snow avalanches.

Glacial–erosional geomorphology

The obvious features produced by alpine glacial erosion of the Tablelands are the major valleys bounding the northeast and southwest flanks and a few tributary cirques and short valleys (Figs. 2, 4). Trout River Gulch is a classic U-shaped glaciated valley 500–800 m deep (Figs. 2, 5), but it lacks a conventional beginning and ending: it is beheaded on the east by the South Arm of Bonne Bay (a fjord), and its downstream end (on the west) is cut off by the mouth of Trout River Pond Valley. Neither anomaly need have prevented a valley glacier from flowing westward through Trout River Gulch in Late Wisconsinan time, but the crosscutting relations suggest that an ancestral Trout River Gulch was left stranded at a relatively high elevation as fluvial and (or) glacial erosion cut down across either end of the gulch.

Most of the bottom of Trout River Pond Valley is covered by water, but dry portions at either end of the valley, and at the narrows joining Trout River Small Pond and Trout River Big Pond (Fig. 2), reveal another U-shaped glacial valley.

In contrast, the east and southeast sides of the Tablelands surface are bounded by a broad, irregular, hummocky bedrock surface, on the order of 300 m lower, that reveals no obvious glacial–erosional character (Fig. 2). This surface at least in part was cut by streams draining northward to Bonne Bay and southward to Trout River Pond Valley, but it may have been modified by glacial erosion in Late Wisconsinan and earlier times.

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Table 3. ^{36}Cl ages and sample data for Trout River Gulch samples.

Sample No.	Sample ratio (^{36}Cl per 10^{15} Cl)	Cl (ppm)	S_t	$\Lambda_{f,e}$ (g/cm^2)	Age (ka) Zero erosion	Erosion rate 1 mm/ka	Erosion rate 5 mm/ka
NF-02-PBC-201	78.2±4.1	372	0.96	173	20.4±1.1	17.3±0.9	13.1±0.7
NF-02-PBC-202	75.8±4.0	596	0.96	174	20.9±1.2	17.8±1.0	13.5±0.8
NF-02-PBC-203	72.6±3.8	329	0.98	172	22.4±1.3	19.0±1.1	14.3±0.9
NF-02-PBC-204	81.2±4.3	308	0.99	171	25.2±1.6	21.0±1.4	15.7±1.0
NF-02-PBC-205	61.4±3.2	153	0.99	341	17.8±3.0	15.6±2.6	12.1±2.0
NF-02-PEN-206	63.1±3.3	435	0.99	168	20.2±2.2	17.3±1.9	13.1±1.5
NF-02-PEN-207	60.2±3.2	275	0.98	172	17.1±1.7	14.9±1.5	11.6±1.2
NF-02-PEN-208	56.0±2.9	335	0.96	174	14.2±2.8	12.6±2.5	9.9±2.0
NF-02-PEN-209	nd	253	—	—	—	—	—

Note: Sample NF-02-PEN-209 is a shielded bedrock sample, so no exposure age was calculated. nd, below detection; S_t, topographic shielding factor; $\Lambda_{f,e}$, effective attenuation length for fast neutrons.

The north rim of the Tablelands, overlooking Trout River Gulch, is dissected by a few glacially eroded reentrants. There are two very shallow cirques, an unnamed one immediately west of the mouth of Winterhouse Gulch, and the Devil's Punchbowl (Fig. 2). Both are perched above the floor of Trout River Gulch. The east fork of Dry Brook (Fig. 3) heads in an arcuate, steep-walled amphitheatre fairly referred to as a cirque, even though a flat floor is not apparent. Winterhouse Gulch is headed by a cirque and is long enough (~2.5 km) to be regarded as a glaciated valley tributary to Trout River Gulch (Fig. 4). Southwest Gulch (Fig. 2), drained by Shoal Brook at the east end of the Tablelands, is a small glaciated valley in the transition zone to the lower hummocky bedrock surface. All tributaries other than Southwest Gulch, including the west fork of Dry Brook, have cut V-shaped fluvial gorges into the north flank of the Tablelands. The cirque headwalls of Winterhouse Brook and the east fork of Dry Brook are incised by fluvial gorges, suggesting that the fluvial incision post-dates the glacial erosion.

Somewhat similar embayments indent the southwest rim of the Tablelands, adjacent to Trout River Pond Valley. The largest is a ~2.5 km long valley incised by Fox Point Brook (Fig. 2). Though not notably U-shaped in cross section, steep bedrock walls rise above talus slopes, and the fluvial gorge at the head of the valley cuts through an arcuate headwall; the valley appears to have been glacially eroded in the past but has been much modified and partly infilled with fluvial and colluvial sediment. Similarly, Long Point Brook (Fig. 2) appears to be a cirque modified by fluvial erosion and talus accumulation, but gorges of Grassy Point Brook, Rocky Brook (Fig. 2), and other, unnamed brooks on this side of the plateau appear to be strictly fluvial features.

Bedrock walls rising above both fluvial and glacial incisions meet the plateau surface at sharp angles. There are no large- or small-scale glacial erosional features on the Tablelands surface itself described in previous literature or observed by us.

Glacial and periglacial deposits

The deposits in and adjacent to the Tablelands consist in part of a rather flat topped grey diamict in the bottoms of some of the tributary valleys, capped by a thin veneer of an-

Fig. 5. Irregular debris bench (IDB) looking eastward along Trout River Gulch from Dry Brook. The bench is discontinuous and in part consists of rock-glacierized debris.

gular colluvial debris. Resting on the diamict in the tributary valleys are rock glaciers. Rock-glacierized debris along the northeast flank of the Tablelands constitutes the "lateral moraine" of previous literature, here termed "irregular debris bench." In the Devil's Punchbowl cirque there is no grey diamict, but there is a cirque moraine, and below it two broad lobate deposits. On the north side of Trout River Gulch is an exposure of brown till. These all are discussed in detail in the following sections. The grey diamict, overlying colluvial diamict, rock glaciers, and irregular debris bench occur in association with each other, and stratigraphic relationships between these units are shown in Fig. 6.

On the Tablelands surface itself, we found no evidence, on airphotos or in the field, of glacial landforms or deposits.

Lateral moraine

Described in all of the relevant works of Grant (1987, 1989*a*, 1989*b*), all the guidebook articles, Berger et al. (1992), and Brookes (1993) is a lateral moraine on the south side of Trout River Gulch. This "moraine" is an irregular, discontinuous, bench-like form, declining in altitude seaward,

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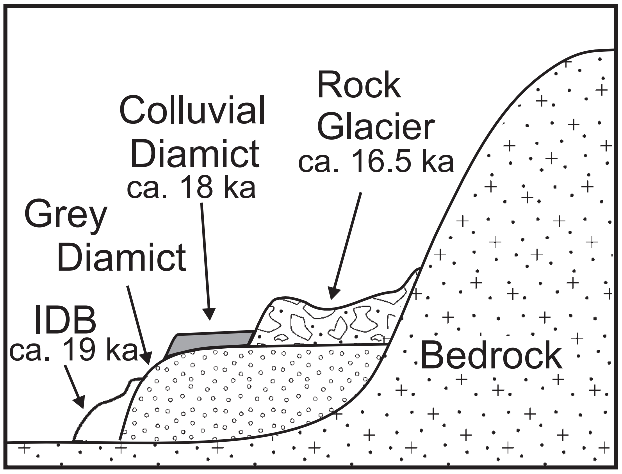
Fig. 6. Cross-sectional cartoon showing typical stratigraphic relationships between bedrock, grey diamict, irregular debris bench (IDB), rock glaciers, and colluvial diamict.



whose sags and depressions are attributed to creep of interstitial ice, i.e., rock glacierization (Grant 1987; Brookes 1993). Part of this feature is labeled "Solifluction lobe" in a photograph in Berger et al.

There are indeed bench-like forms along parts of the valley wall, ranging from 50 to 100 m above the valley bottom. Some of the segments are 50–70 m broad, irregular, in part hummocky terraces sloping gently toward the valley; in a few cases, small, local depressions resembling sinkholes are bounded by ridge crests (Fig. 5). These forms most likely did not originate as a lateral moraine, however, for the following reasons. (1) The "moraine remnants" are composed almost entirely of angular (0.17–0.25 on the scale of Lewis and McConchie 1994), unpolished blocks of cobble to boulder size, typical of rockfall debris, and identical to debris littering the bedrock slopes above. There are no polished and (or) abraded clasts typical of till of large valley glaciers. Furthermore, the clasts along parts of the "moraine" are free of interstitial fines, typical of rockfall debris but not typical of lateral-moraine till. (2) There are rare quartzite and other exotic clasts in some of the debris, indicating some input from a former valley glacier, but at least 99% of the clasts are peridotite like that on the slopes above, suggesting that almost all the debris is from the slopes above. (3) West of Dry Brook (Fig. 3) there is a segment of bench, apparently undisturbed by rock glacierization, that rises in elevation to the west, whereas the valley bottom drops to the west. (4) West of Berry Barren Brook (Fig. 3), upslope of the parking lot of the Green Gardens Trail (Fig. 2), there are *three* benches, and colluvial cover is thin enough to reveal that the benches are bedrock controlled. (5) Still farther west, where the lithology is gabbro instead of highly fractured peridotite, much less talus is produced and, probably not coincidently, there is no hint of lateral moraine. (6) There are no equivalent moraine remnants on the north side of Trout River Gulch.

Although we cannot outright preclude that the discontinuous slope deposit along the base of the north flank of the Tablelands comprises fragments of lateral moraine, we favour an alterative interpretation that the landform is in part glacial drift, in part colluvium reworked in places by rock glacierization, and in part bedrock-controlled steps. The local depressions along the bench west of Dry Brook appear to be due to collapse of debris into void space created by melting ice (see the section titled Rock glaciers). In the same area, there are debris-flow levees and rockfall hummocks overlying the bench whose boulders are noticeably less weathered than those of the bench. This suggests that the bench debris collected mainly during a past episode of debris production–accumulation. Although a strict lateral moraine interpretation seems untenable, it is quite possible that much of the angular rock, though derived from slopes above, at one time crowded the margin of a decaying valley glacier in Trout River Gulch. Some of it may have been debris that covered marginal ice, which was protected as clean ice in the middle of the valley ablated. Slow, delayed collapse of debris-rich, ice-saturated margins could help explain the rock glacierization.

Trout River Pond Valley end moraine

Brookes (1993) mapped an "end moraine" adjacent to ice-marginal sediments near the mouth of Trout River Pond Valley, which he thought was deposited by a piedmont glacier extending out of that valley. The surficial geology map in Berger et al. (1992) shows no such moraine, and we are unable to find such a moraine either in the field or on airphotos.

Cirque moraines in the Devil's Punchbowl and Winterhouse Gulch

According to Grant (1987), and repeated in later guidebooks, small cirque moraines occur at the head of Winterhouse Gulch and the Devil's Punchbowl, a small tributary valley and cirque, respectively, on the south side of Trout River Gulch (Fig. 2). Grant's report states that "The younger one is distinctly less oxidized than the older" (p. 53). It is not clear whether this means that there are two moraines at each of the sites or that a moraine at one of the sites is less oxidized than that at the other site. The surficial geology map in Berger et al. (1992) shows a cirque moraine in Devil's Punchbowl but not in Winterhouse Gulch. Brookes (1993, p. 73) described the moraine in Devil's Punchbowl and noted that a "very small, fresh moraine stands at 550 asl below the central headwall of Winterhouse Brook trough." Grant described the cirque moraines as probably of Neoglacial age, Berger et al. claimed that the Devil's Punchbowl moraine debris is "virtually unweathered" and suggested a possible Little Ice Age (LIA) origin, and Brookes suggested a "Late Wisconsinan" age for the Punchbowl moraine and an LIA age for the Winterhouse Brook moraine. We see no evidence for a moraine in Winterhouse Gulch, either in the field or in airphotos. However, the Punchbowl moraine is very conspicuous.

The Devil's Punchbowl is a shallow cirque approximately 750 m wide, incised 200 m into the north flank of the Tablelands plateau (Fig. 3). The cirque headwall is very fractured. Most of the bedrock surface is weathered to the characteristic brown colour, but on the headwall are scattered zones of irregularly fractured rock with the dark grey–green colour of unweathered peridotite. These zones generally are situated above debris-flow tracks on the apron below and represent sites of mass movement. The base of the headwall is mantled by colluvium; gully exposures reveal at least a 3.5 m thickness of partly matrix supported and partly clast sup-

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ported colluvial diamict. The surface of this sediment is marked by many debris-flow levees and lobes. The considerable accumulation of colluvium in the cirque suggests there was no recent occupation of the cirque by active ice, although there are no data with which to construct a quantitative argument.

The moraine (Fig. 7) is a curved ridge 20–25 m high on its distal flank and 0–8 m high on its proximal flank, depending on the volume of more recent debris-flow sediment that has piled up on the latter. The easterly part of the ridge has a simple, single crest, and the westerly portion is broader and more irregular, with more than one simple crest but not clear multiple crests. The middle part of the ridge is buried by debris-flow sediment.

The ridge is composed of peridotite clasts of all sizes up to 5 m long. Much of the surface is bouldery without interstitial fines, but the western half features patches of boulder-studded sand and silt, even bearing some krumholtz and a few shrubs. The boulders are mostly angular to subangular (0.17–0.35 on the scale of Lewis and McConchie 1994), with a few subround (0.35–0.49) exceptions. The boulder surfaces in general appear weathered, with shallow gnamma, in situ fracturing, and weathering along fractures. Weathering rinds on the boulders are inconsistent: some surfaces have a ~1 mm thick brown rind over fresh rock; some have a ~1 mm brown rind over a usually 4–5 mm (but as much as 10 mm) thick, less well-developed rind over fresh rock; and some have the lesser, thicker rind but not the stronger, thinner rind. Rind thicknesses and inconsistencies are similar on the older lobate deposits below the cirque (see later in the paper) and on the Tablelands plateau itself, implying that rind thicknesses exist in some kind of steady state: most likely removal of mass proceeds at about the same rate as rind formation, so rinds do not thicken with time.

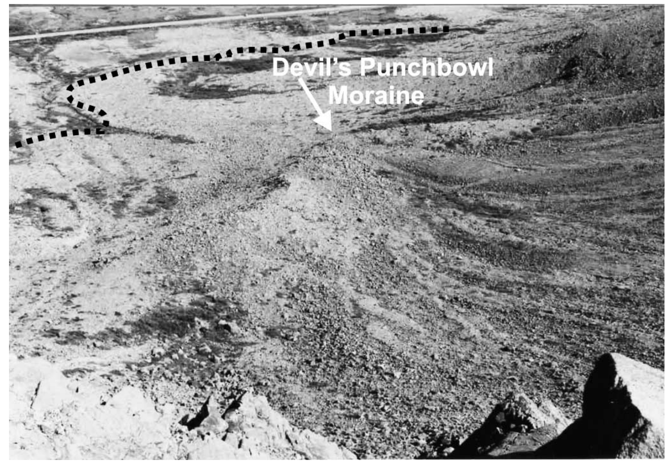
We confidently interpret the ridge to be a moraine, as have many before us. It is arcuate in plan view and fits appropriately with the plan of the cirque. It is not a protalus rampart because (*i*) it bulges away from the cirque headwall instead of paralleling the base of the headwall, (*ii*) it contains a considerable volume of fines, and (*iii*) the requisite snowbank would have a very low gradient.

Other cirque moraines?

We do not recognize any other cirque moraines on or about the Tablelands, outside of the Devil's Punchbowl. This distribution of moraines is indeed problematic: if a cirque moraine remains well preserved in the Punchbowl, there is no apparent reason why, for example, the nearly identical cirque just west of the mouth of Winterhouse Gulch (Fig. 2) does not contain a moraine. Furthermore, currently (judging from photographs taken in winter) there is no more than average snow accumulation in the Punchbowl relative to that in the surrounding slopes, and in June more snow remains on the west side of the Punchbowl than against the headwall above the moraine.

Unpredictable cirque moraine distribution prevails elsewhere in the Long Range Mountains: rare cirque moraines occur, for example, at the north base of Gros Morne Mountain (the highest mountain in the park; Fig. 1), but most cirques are without moraines.

Fig. 7. Devil's Punchbowl moraine and lobate deposits (broken line) below, as seen from the rim of the Tablelands.



Devil's Punchbowl cirque lobate deposits

In the lower reaches of the Devil's Punchbowl cirque, below the cirque moraine, the irregular bench interpreted in guidebook articles to be a lateral moraine actually consists of two adjacent lobes of debris with surface depressions similar to those of hummocky moraine (Fig. 7). Unlike the cirque moraine, the lobes contain a few percent or less of exotic clasts, mainly quartzite, indicating that a valley glacier in Trout River Gulch once added exotic material to the local peridotite debris. However, the lobate fronts of these deposits clearly indicate that the debris flowed from the area of the cirque to the margin of the gulch, such that any original contributions of till have been reworked and partially covered. We speculate that the lobes are rock-glacierized debris that was originally situated in the lower cirque and adjacent valley walls. Based on morphology or rind thickness alone, there is no way to establish relative ages of the lobate deposits and the cirque moraine; it is possible that the lower debris was flowing while a cirque glacier higher up was depositing the moraine.

Rock glaciers

Rock glaciers in the Penman's Brook (Dry Brook in this paper) and Winterhouse Brook drainages were noted by Berger et al. (1992) and Brookes (1993); Brookes also mapped them in the Fox Point Brook drainage, on the south side of the Tablelands (Fig. 2). The rock glaciers were regarded by these authors to be talus accumulations activated by interstitial ice, and Brookes also noted the presence of protalus ramparts in the Fox Point Brook drainage.

In general, rock glaciers on the flanks of the Tablelands are of the valley-wall variety, what Whalley and Azizi (1994) would call "protalus lobes" (Fig. 8). None show any signs of recent activity (Wahrhaftig and Cox 1959). Most are found within tributary valleys incised back from Trout River Gulch and Trout River Pond Valley, where they rest on the top surface of the grey diamict discussed in the next section (Fig. 6). Although some valley-side hummocks near the head of Winterhouse Brook may be of rock-glacier origin, the best forms are in the valleys of Fox Point Brook and Dry Brook.

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Fig. 8. Small cliff-base rock glacier (RG) resting on the grey diamict near the mouth of Dry Brook.

Along the east side of Fox Point Brook, an irregular but continuous line of small, west-facing rock glacier fronts rises as high as 40 m above the surface of the grey diamict (Fig. 9; see later in the paper). Behind the frontal crest are scattered irregular depressions as deep as 25 m. The rock glaciers are composed of weathered peridotite boulders up to several metres long, some openwork and some interstitially filled with fines and bearing small patches of juniper krumholtz. No clasts of exotic lithology were found.

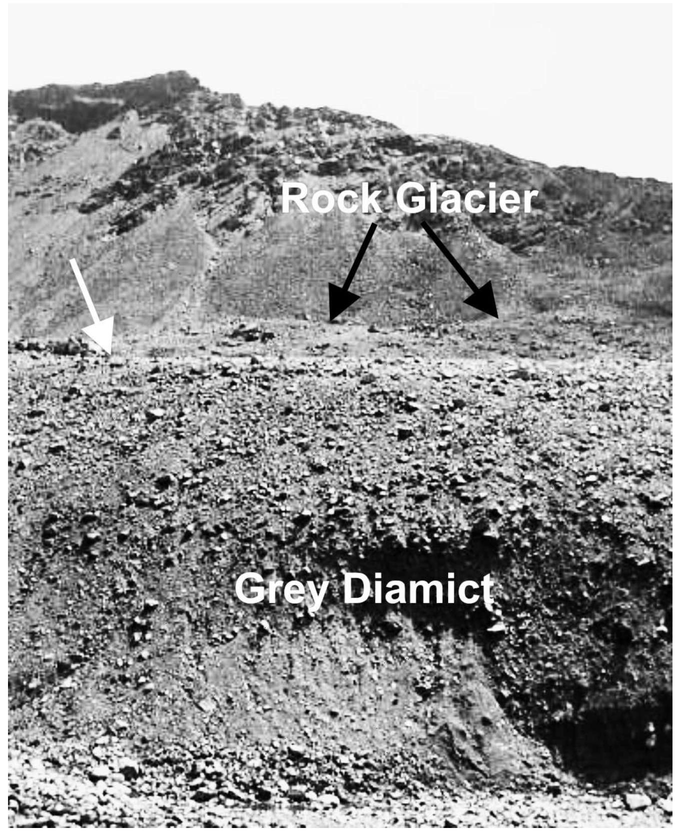
On the west side of this small valley there are remnants of similar valley-wall rock glaciers perched on the slope, but incision of the brook appears to have removed much of the rock glacier mass. There probably is grey diamict beneath these rock glaciers, but there remains no geomorphic expression of it; incision has created a continuous steep slope between rock glacier remains and the watercourse, without exposures.

In the valley of Dry Brook, valley-wall rock glaciers are found on both sides of the east fork and on the west side of the west fork (Fig. 3). In the one case where an underlying fill surface is preserved, the rock glacier front rises approximately 50 m above the fill (Fig. 8). The deepest depression behind a frontal crest is approximately 15 m deep. Depressions tend to be cone- or wedge-shaped rather than rounded, even where not partly filled with later debris-flow material. The composition and size of blocks, variable content of interstitial fines, and lack of exotic lithologies are the same as described along Fox Point Brook.

The line of rock glaciers immediately east of the confluence of the two forks of Dry Brook remains intact, and remnants of the underlying tributary-fill surface are preserved, but farther up the east fork, and along the west fork, stream incision appears to have removed variable amounts of the original frontal portions of the rock glaciers. A pit was dug in one of these eroded deposits on the west side of the east fork, revealing clast-supported angular (0.17–0.25 on the scale of Lewis and McConchie 1994) blocks with an interstitial fill of angular coarse sand, granules, and fine gravel. The sediment contains very little silt and clay, which distinguishes it from the underlying tributary-valley fill of grey diamict.

In the very brief descriptions of Berger et al. (1992) and Brookes (1993), the rock glaciers were considered to be talus activated by interstitial ice. However, the sharp frontal

Fig. 9. Incised fill of the grey diamict in the valley of Fox Point Brook (which runs at the bottom of the frame). Rock glaciers rest on the nearly flat surface (white arrow) of the diamict.



crests and closed depressions behind the crests indicate that there were significant volumes of ice in the cores of these rock glaciers when they formed. If the ice had been truly interstitial, the rock glaciers would have been only marginally matrix supported, and little or no deflation of the top surface would have occurred when the ice ablated (Wahrhaftig and Cox 1959; Barsch 1987; Clark et al. 1998). Absence of bedrock exposures between the frontal crests and the talus aprons fed from the cliffs behind suggest that the ice masses that formed these deposits were debris-covered. The genesis of ice-cored, valley-wall (as opposed to cirque) rock glaciers is not well understood (e.g., Whalley and Azizi 1994; Clark et al. 1998). We suggest that these features (like similar ones studied in the Sierra Nevada; Clark et al. 1994) formed as a result of abundant snow and debris avalanches during the spring and early summer, at a time when snow accumulation was more substantial than it is now. The freeze–thaw process is most active in spring, and early avalanches would contain more snow than rock debris. Seasonally later avalanches would be more debris-rich, and late-stage debris would act to protect the buried snow–ice from earlier in the year (or from heavy years). Such a process would form nearly debris-free layers within the talus, which would provide the means for the combined ice and debris mass to flow downhill. In a later time of greater warmth and less snow, the buried ice would melt, creating the depressions.

The rock glaciers are perched up on the sides of their respective valleys, higher than they would be if their base lev-

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els had been present stream levels. We hypothesize that the rock glaciers were deposited on sedimentary fills extending across the bottoms of each of the tributary valleys; near the head of the east fork of Dry Brook, rock glaciers from both east and west sides of the cirque coalesced on top of the fill. Subsequently, streams incised through the fills, often into bedrock below, leading to erosion of some of the rock glacier toes. However, *some* incision preceded rock glacierization, because the front of one of the rock glaciers in the east fork of Dry Brook cascades over an erosional depression in the fill.

Grey diamict

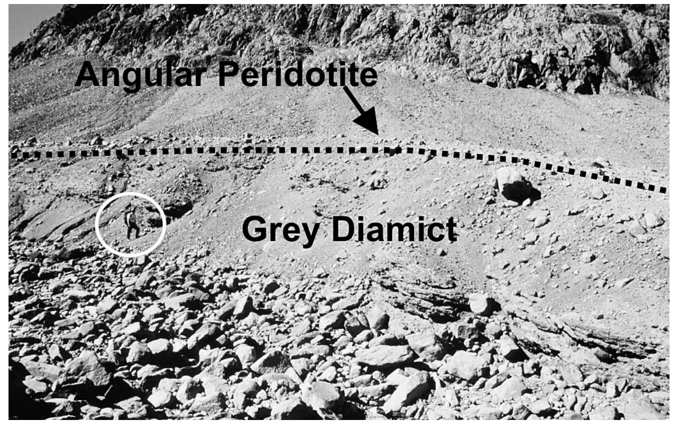
A black to grey, peridotitic, well-consolidated, matrix-supported diamict, herein referred to as the grey diamict, underlies colluvium, rock-glacierized debris, and talus in all tributary valleys investigated and also extends into the main trunk valley (Trout River Gulch) at Dry Brook and Winterhouse Gulch (Figs. 6, 10, 11). It comprises the sedimentary fill noted previously. The only previous reference to any of this deposit is by Brookes (1993), who described a stratified, gently dipping diamict near the head of Winterhouse Brook (Fig. 10) and suggested it may be a flow till at the “Late Wisconsinan glacial limit” in that small valley. The diamict occurs as a thick (greater than 40 m in some cases), rather flat-topped blanket in the bottoms of tributary valleys but does not continue up the sides of those valleys. The relatively flat top is expressed best at Fox Point Brook (Fig. 9), where the top slopes gently southwestward (downvalley), until cut off by incision of the brook. The surface projects out into Trout River Pond Valley, high above Trout River Pond, suggesting that the diamict is a fill that was dammed against a glacier in the valley. On the north side of the Tablelands, bodies of the diamict are thinner, and top surfaces less distinct.

The limited exposures available indicate that almost all clasts are peridotite; there are rare (<1%) exotic clasts of quartzite, argillite, and limestone. The sediment (*i*) in general contains a rather complete spectrum of grain sizes (there is no great distinction between “clasts” and “matrix”), (*ii*) in many cases exhibits a high degree of fissility with occasional polished surfaces (fissility in Winterhouse Gulch dips 10°–15° upvalley), and (*iii*) in some cases exhibits crude stratification that includes discontinuous silt–clay lenses and (or) openwork beds and (or) very poorly graded beds (beds where present tend to dip 10°–25° downvalley). Some minor folding and apparent shearing are evident where there is stratification. Bedding and openwork structure are more common in upper parts of the diamict, where there is also some valley-parallel, long-axis orientation of elongate clasts. Sections in basal contact with bedrock contain clay-rich diamict in which clasts are commonly subangular to subrounded (0.25–0.49 on the scale of Lewis and McConchie 1994), facetted, and striated. Travertine cementation is common at all sites and especially well developed at the bases of the sections; the cement precluded detailed fabric analysis.

Dipping, crudely sorted and stratified gravel alternates with massive, matrix-rich diamict along Fox Point Brook.

We interpret the diamict to be partly (particularly the lower part) till; consolidation and fissility suggest lodgement till. The upper part shows no classic lake sediments (e.g., laminated silt), but there is much evidence of water working or washing (clay stringers, openwork beds, crude sorting), suggestive of water-saturated debris flows in a proglacial environment. The complexity of this sediment suggests a complex depositional history that may have involved cycles of lodgement, deformation, and water working at the top or base of ice, consistent with deposition in a wet-based, high-gradient environment (Lawson 1979; Ham and Mickelson 1994; Spooner 1994). Certainly many of the features described are consistent with both lodgement and melt-out processes; however, we saw no direct evidence for two distinct tills. The rather flat top of the diamict at Fox Point Brook and its projection into Trout River Pond Valley suggest the sediment may have been ponded against a glacier occupying the valley at the time.

Fig. 10. Grey diamict in Winterhouse Gulch, exposed by postglacial stream incision and overlain by a veneer of angular peridotite. Broken line represents approximate contact. Part of exposure to right of person (circle) shown in Fig. 11. Modern stream deposits constitute foreground rubble.



Diamict veneer

A thin veneer of very angular to angular (0.12–0.25 on the scale of Lewis and McConchie 1994) cobbles and boulders overlies the grey diamict (Fig. 6). This unit is composed almost entirely of peridotite, but occasional quartzite erratics were noted at locations in Trout River Gulch. The angular nature of the clasts suggests an origin in the rubble of bedrock slopes and (or) rock glaciers, but the process by which such rubble is spread out over the grey diamict is not clear.

Brown diamict

A light to dark brown polymictic, well-consolidated, matrix-supported, coarse diamict crops out in cutbanks along Trout River near the Green Garden Trail, on the north side of Trout River Gulch (Fig. 2). Clasts are subrounded to well rounded with long axes to 40 cm, arranged in a weak fabric parallel to the valley axis, and rarely striated. Clasts consist of peridotite and gabbro, with a few percent gneiss and quartzite. The matrix is mainly sand and silt, cemented by calcite, and exhibits a wavy, subhorizontal fissility and some minor, local sorting and stratification. We interpret this diamict to be till, presumably left by a valley glacier in Trout River Gulch; it conceivably could be the partial equivalent of the

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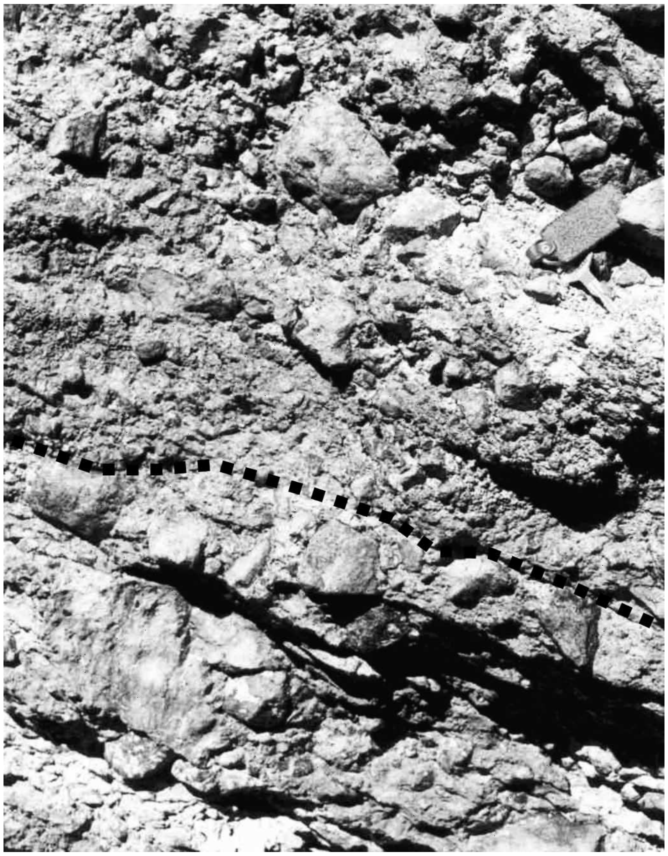
Fig. 11. Closeup of grey diamict. Crude stratification (broken line) in part of the deposit dips to the right. Note key fob in upper right for scale.



grey diamict, coloured differently by the different source rocks on the north side of the valley.

Chronology

The concentrations of ^{36}Cl were measured in eight boulders and one geological blank sample. The geological blank, a shielded bedrock sample of the peridotite (with depth >5 m), yielded no measurable ^{36}Cl, so noncosmogenic ^{36}Cl is likely to be negligible in all samples. This is consistent with the low concentrations of targeted trace elements that could lead to radiogenic or nucleogenic ^{36}Cl (Table 2).

If the concentrations are interpreted as simple continuous exposure ages, the ages range from 14.2 to 25.2 ka (Table 3). These are apparent ages, not corrected for snow cover or erosion. Uncertainties in Table 3 for individual ages represent the 1σ precision in AMS measurement. There are two pairs of samples from the same geomorphic position (samples NF-02-PBC-201 and NF-02-PBC-202 are from the western side of Devil's Punchbowl moraine, and samples NF-02-PBC-203 and NF-02-PBC-204 are from the lobate deposits in front of the same moraine). The ages of the twins are within 1σ uncertainty for each pair, which may indicate that differential environmental adjustments for shielding were reasonable and that the sample ages of different boulders are not significantly influenced by differential erosion. The coefficient of variation for AMS analysis of a chemical duplicate of sample NF-02-PBC-202 was 3.2%, also within the 1σ uncertainty for each measurement. When comparing with ages determined from other chronometers, such as calibrated radiocarbon ages, the uncertainty in these exposure ages should consider all significant random and systematic error, which may approach 15% at 1σ (Phillips et al. 1996; Gosse and Phillips 2001).

Exposure ages are influenced significantly by erosion of the boulder surfaces (Fig. 12). None of the boulder surfaces sampled could be assumed to be original. In the region, boulders on the Devil's Punchbowl cirque moraine show weathering in the form of rounded edges, gnammas, and rillen, particularly when the coarse pyroxene cumulates were exposed. On some boulders, veins of more resistant composition (e.g., fine serpentinite) protrude more than 1 cm above the surface. For at least two of the eight boulders, the erosion appears to be less on the upper surfaces than on the boulder sides; nevertheless, some boulder tops exhibited indications (e.g., grain relief) of 1 cm erosion. Less than 50 km to the north on Big Level, coarse Precambrian granitic gneiss at 800 m is eroding at rates of ~1 mm/ka, as indicated by the 20 ka exposure age of striated quartz veins that protrude <20 mm above the gneiss tor surface (Gosse et al. 1993). Furthermore, we did not observe sombrero-shaped boulders (e.g., Zimmerman et al. 1994) anywhere in the study area. Sombrero-shaped boulders are formed when subaerial erosion creates a brim at ground level. The brim width is proportional to erosional loss. The lack of sombrero rims may indicate that long-term average surface erosion rates probably are less than 5 mm/ka because over 15 ka a conspicuous 75 mm would have eroded from the portion of the boulder above ground (60 mm brims were easily recognized by Zimmerman et al. 1994). The actual amount of thickness that would be lost from a boulder surface averaged over all eight samples is shown by the thick line in Fig. 12. An erosion rate of 5.0 mm/ka (i.e., 63 mm eroded based on the average of all eight samples) is considered an upper limit for the specific boulders sampled, considering the previous observations and our sampling strategy, which favours boulders with low erosion. Ages in Table 3 are calculated for constant erosion rates of 0, 1.0, and 5.0 mm/ka (Fig. 13). Note that doubling the erosion rate to 10 mm/ka will not significantly decrease the age. Although we believe the erosion rates of the selected boulders are much less than 5 mm/ka, because we cannot preclude that the boulder erosion rates were not as high as 5 mm/ka we use the range of ages derived from 1 and 5 mm/ka erosion rates in our interpretation. We do not consider the 0 and 5 mm/ka erosion rate ages to indicate an upper and lower limit or uncertainty range of the ages but provide them to demonstrate the sensitivity of these ages to erosion and the possible bounds of the exposure ages.

The partial shielding of fast neutrons and capture of thermal neutrons by snow can cause unadjusted ages to underestimate the actual duration of boulder exposure. Estimations of the influence of snow cover on ^{36}Cl require knowledge of snow density, thickness, and duration, which is unavailable at our sampling site. However, based on observations of snow cover since the winter of 2003–2004, the effect of snow is probably <10%. However, snow cover may have been more or less significant in the past. There is a weak positive trend in age with boulder height, which may reflect the effect of snow cover, but there are too few measurements

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Fig. 12. Chlorine-36 exposure ages for all eight samples (last three digits of the sample numbers are given in the legend) adjusted for boulder surface erosion. Lines show the grouping of the samples according to landform (cf. Fig. 13). Total thickness of rock that would have been removed from the surface sampled is represented by the thick line, based on an average of all eight boulders.

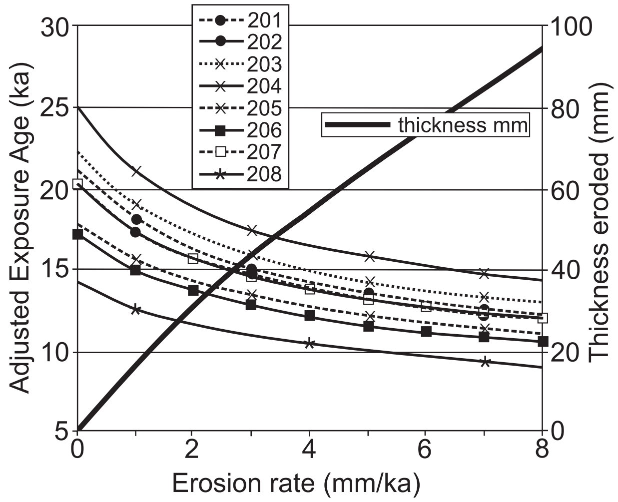
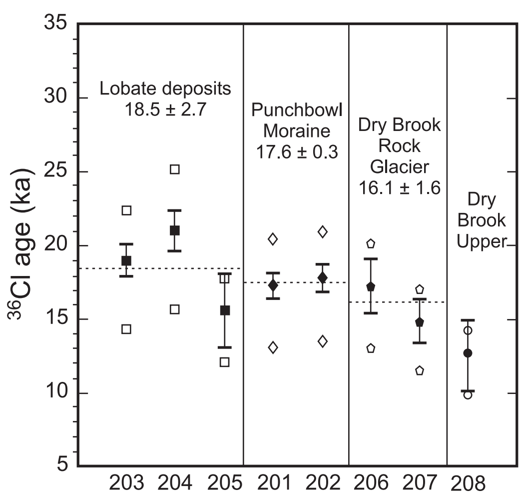


Fig. 13. Cosmogenic ^{36}Cl ages for Trout River Gulch boulder samples (the last three digits of the sample numbers are given along the *x* axis). Ages assuming no erosion are the oldest for each boulder; solid symbols are adjusted for 1 mm/ka erosion; and lowest ages assume 5 mm/ka erosion on each boulder. Error bars represent 2σ uncertainty in AMS precision. Mean age (broken line) for each landform is indicated with standard deviation.



to provide confidence. For instance, sample NF-02-PEN-206 (45 cm above ground level; Table 1) yielded an exposure age that was significantly less than those of the two higher boulders on the lobate deposits (Fig. 13). Based on the available data, we have not adjusted the ages for the effects of snow, and we recognize that for the ^{36}Cl ages in this study, the effect of snow cover is in an opposite sense to the effect of erosion.

It is possible that some ^{36}Cl was present in the boulders before they were deposited at their sample sites. This inheritance would cause the exposure ages to overestimate the age of the landforms. However, considering the close agreement among the pairs and the overall consistency of the ages, and their stratigraphic order, there is no evidence to suggest inheritance is a factor. Our sampling strategy minimized the likelihood of post-depositional movement.

Discussion

Significance of TCN ages

The TCN ages indicate that the rock glaciers and cirque moraine, which at first glance would seem to be (and indeed previously have been interpreted to be) Holocene deposits, are records of Late Pleistocene ice in the small valleys incising the Tablelands. The absolute minimum erosion-adjusted age (requiring an extreme erosion rate of ~30 mm/ka) is >9 ka for all samples (except sample NF-02-PEN-208, which is younger), but this would require erosion of 27 cm, the result of which can be readily recognized and avoided in the field.

Samples NF-02-PBC-201 and NF-02-PBC-202 indicate an age of 17.6–13.3 ka for the Devil’s Punchbowl cirque moraine. Samples NF-02-PBC-205 and NF-02-PEN-206 indicate an age of 16.1–12.6 ka for the Dry Brook rock glacier. These ages are consistent and show that in Late Wisconsinan time, well after the LGM, the flank of Trout River Gulch was the site merely of periglacial activity and minor ice accumulation; ice in the Punchbowl terminated on the flank of the Tablelands rather than merging with a valley glacier in the gulch. Sample NF-02-PEN-208, from another Dry Brook rock glacier, seems anomalously young at 12.6–9.9 ka, but may represent a Younger Dryas age restabilization of ice. The small boulder (60 cm high) may have been affected by snow cover and therefore may represent an underestimate of the actual exposure duration.

The lobate deposits below the Punchbowl moraine are dated by samples NF-02-PBC-203, NF-02-PBC-204, and NF-02-PBC-205 and average 18.5–14.1 ka (ranging from 21.0 to 12.1 ka). Because sample NF-02-PBC-205 was the smallest boulder sampled, at only 45 cm high, we consider it most likely that its anomalously low age underestimates the age of the lobe surface and that the other two samples best represent the age of the lobes (means of 20.0 and 15.0 ka for the 1 and 5 mm/ka erosion rates, respectively). The ages indicate the lobes are similar in age and somewhat older than the cirque moraine. The lobes rest on the bottom of Trout River Gulch and require the valley to be ice free sometime between 20 and 15 ka.

We use the ages of the Punchbowl deposits and the Dry Brook rock glaciers to suggest that all or at least most of the rock glacierization along the flanks of the Tablelands, including that in Fox Point Brook on the south side, is of Late Wisconsinan age.

The boulder that provided sample NF-02-PEN-206, yielding an age of 17.3–13.1 ka, is part of the thin veneer on top of the grey diamict near the mouth of Dry Brook. Although we are not sure by what processes the angular debris in general was spread across the fill top, it is possible that the sampled boulder rolled onto the surface from the adjacent rock glacier, which yielded a similar age. Sample NF-02-PEN-206 provides a minimum age of the grey diamict, but we believe the grey diamict is considerably older than the rock glaciers and angular debris that rest upon it. The diamict is likely a prod-

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uct of subglacial ice that filled at least the tributary valleys of the Tablelands and so must represent an earlier glacial episode than the cold and (or) snowy period that produced the rock glaciers and Devil's Punchbowl moraine.

If future work indicates that erosion rates are greater than we have estimated, then the exposure ages are commensurately younger (to a minimum of 9 ka; see earlier).

Distribution of LGM and later ice

Of the three basic geographic units of the Tablelands area (Trout River Pond Valley, the Tablelands surface, and Trout River Gulch), the gulch we interpret to have been relatively ice-free during post-LGM Late Wisconsinan time. If our preferred erosion rate of 1 mm/ka is used, the resultant 20 ka age of the lobate deposits suggests there was no valley glacier in the gulch five or more millennia after the LGM time favoured by Shaw et al. (2006). In this scenario, the large valley glacier indicated by the gulch's transverse U shape, a large granite erratic on the floor of the valley, and exotic clasts mixed with peridotite on the lower flanks of the Tablelands *may* have existed during the LGM. However, according to the reconstructions of Shaw et al., the ice margin in the Gulf of St. Lawrence remained generally at its maximum extent through 19 cal ka, which provides no rationale for any change in ice volume in the gulch between the LGM and 19 cal ka. The implication is that any large valley glacier in the gulch probably existed prior to Late Wisconsinan time, as postulated by Brookes (1993). With the 5 mm/ka erosion rate, a valley glacier at the LGM is easily accommodated, although the ages of the lobate deposits indicate at least an early deglaciation of the gulch, possibly influenced by the beheaded nature of the upstream end of the gulch.

Trout River Pond Valley, in contrast, was occupied by a valley glacier in very Late Pleistocene time. The evidence comes not from the Tablelands but from the marine delta at the seaward end of Trout River Pond, with ages of 13.8 and 13.2 cal ka (Brookes 1974; Grant 1987). Ice also is thought to have extended into the Gulf of St. Lawrence off the mouth of Bonne Bay at about that time (Shaw 2003), suggesting that in very Late Pleistocene time ice was streaming seaward both north and south of a largely ice-free Trout River Gulch. LGM ice in Trout River Pond Valley may have been the ice that dammed the fill of grey diamict in Fox Point Brook.

As for the plateau itself, we have no direct evidence that it either was or was not ice-covered at the LGM or during the post-LGM Pleistocene. Tors and felsenmeer indicate no glacial erosion, but erratics on the top indicate ice cover at some time. TCN data from other, similar plateau sites in the Gros Morne region suggest there probably was an ice cover (Gosse et al. 2006), but if so the ice must have been thin and cold-based on the Tablelands: there is no geomorphic or depositional evidence of flow from the plateau to the tributary valleys. Steep headwalls in both cirques and tributary valleys, and sharp breaks between headwalls and the plateau surface, suggest that connections between the plateau and the valleys, if any, were not vigorous. The grey diamict offers no evidence of connections between plateau and valleys: the bodies of diamict abut headwalls instead of extending upward toward the plateau.

The Shaw et al. (2006) synthesis of deglaciation of Atlantic Canada uses mainly marine evidence, supported by TCN exposure ages on boulders on St. John's Highlands, to the north of Gros Morne National Park (Gosse et al. 2006). According to their estimates, the west coast of Newfoundland and adjacent sea were ice-covered through ca. 16.8 cal ka, but by ca. 14.8 cal ka the ice margin was beginning to work eastward from an exposed west coast. Our results using our preferred rate of erosion are inconsistent with this model and suggest that as early as 20 ka there was not a continuous ice cover over all of the western coast. The younger ages using 5 mm/ka of erosion are indeed consistent with the model, however. Resolution depends, as usual, on further work.

Conclusions

(1) Alpine glacial deposits on the flanks of the Tablelands constrain the timing of valley glaciation adjacent to the Tablelands.
(2) We are unable to confirm the existence of (*i*) an original lateral moraine along the south side of Trout River Gulch, although we agree exotic clasts were introduced by a valley glacier at some point; (*ii*) an end moraine near the mouth of Trout River Pond Valley; and (*iii*) a cirque moraine(s) at the head of Winterhouse Gulch.
(3) There is a cirque moraine in the Devil's Punchbowl, but moraines do not occur in other cirques for reasons unknown.
(4) Rock glaciers probably began as masses of avalanched snow interleaved with talus.
(5) Rock glaciers and the Devil's Punchbowl cirque moraine formed in post-late glacial maximum (LGM) Late Wisconsinan time.
(6) A diamict filling the bottoms of valleys tributary to Trout River Gulch and Trout River Pond Valley contains till and debris-flow deposits and predates the rock glaciers and cirque moraine.
(7) Trout River Pond Valley contained an active valley glacier until very late in Pleistocene time, but Trout River Gulch at a minimum was deglaciated unexpectedly early and, if the erosion rate of 1 mm/ka is correct, likely was not actively glaciated during the LGM.
(8) Any Late Wisconsinan ice that covered the surface of the Tablelands must have been thin, as there is no evidence of connections between surface ice and ice on the flanks.
(9) Our conclusions may or may not be consistent with a regional synthesis generated mainly from marine evidence, depending on what erosion rate is assumed.

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