VOLCANICLASTIC SEDIMENTATION IN LACUSTRINE SETTINGS

DEDICATION

Mario Martín Mazzoni, 1943–1999

Dr Mario Mazzoni died on Friday 1 October 1999 of a heart attack, in Quilmes, Argentina. Mario was 56 years old. He received his PhD from the Universidad Nacional de La Plata (UNLP), Argentina, and there, together with a few associates, established the Centro de Investigaciones Geologicas (Centre for Geological Investigations). He was also a senior scientist with CONICET (Consejo Nacional de Investigacion Cientifica y Tecnica: National Consortium for Scientific and Technical Investigation) and Professor of Geology at UNLP.

Mario took his family to Santa Barbara, California, in 1982 to undertake a post-graduate fellowship there with R. V. Fisher. He began a long-term collaboration with R. V. at that time, and in 1988 came back to Santa Barbara to join a group of graduate students (including the editors of this volume) travelling around the western USA looking at calderas and stratovolcanoes. Mario was the premier authority on volcaniclastic rocks in Argentina, most recently beginning investigations of the Quaternary Copahue volcano and a proposed Caviahue caldera. Mario loved to travel, and welcomed colleagues from around the world to Argentina.

Mario was a kind, gentle, generous person and is sorely missed by those of us who were fortunate enough to be able to work with him, or just cross paths.

SPECIAL PUBLICATION NUMBER 30 OF THE
INTERNATIONAL ASSOCIATION OF SEDIMENTOLOGISTS

Volcaniclastic Sedimentation in Lacustrine Settings

EDITED BY

JAMES D. L. WHITE AND NANCY R. RIGGS

Blackwell
Science

Editorial Offices:
Osney Mead, Oxford OX2 0EL
25 John Street, London WC1N 2BS
23 Ainslie Place, Edinburgh EH3 6AJ
350 Main Street, Malden
MA 02148-5018, USA
54 University Street, Carlton
Victoria 3053, Australia
10, rue Casimir Delavigne
75006 Paris, France

Other Editorial Offices:
Blackwell Wissenschafts-Verlag GmbH
Kurfürstendamm 57
10707 Berlin, Germany

Blackwell Science KK
MG Kodenmacho Building
7–10 Kodenmacho Nihombashi
Chuo-ku, Tokyo 104, Japan

Iowa State University Press
A Blackwell Science Company
2121 S. State Avenue
Ames, Iowa 50014–8300, USA

First published 2001

Set by Graphicraft Limited, Hong Kong
Printed and bound at the Alden Press, Oxford and Northampton

DISTRIBUTORS

Marston Book Services Ltd
PO Box 269
Abingdon, Oxon OX14 4YN
(*Orders*: Tel: 01235 465500
Fax: 01235 465555)

USA
Blackwell Science, Inc.
Commerce Place
350 Main Street
Malden, MA 02148-5018
(*Orders*: Tel: 800 759 6102
781 388 8250
Fax: 781 388 8255)

Canada
Login Brothers Book Company
324 Saulteaux Crescent
Winnipeg, Manitoba R3J 3T2
(*Orders*: Tel: 204 837 2987)

Australia
Blackwell Science Pty Ltd
54 University Street
Carlton, Victoria 3053
(*Orders*: Tel: 3 9347 0300
Fax: 3 9347 5001)

A catalogue record for this title is available from the British Library

ISBN 0-632-05847-1

Library of Congress
Cataloging-in-publication Data
Volcaniclastic sedimentation in lacustrine settings / edited by Nancy R. Riggs and James D. L. White.
p. cm.
ISBN 0-632-05847-1
1. Volcanic ash, tuff, etc. 2. Sedimentation and deposition. 3. Lake sediments. I. Riggs, Nancy R. II. White, James D. L.

QE461.V632 2001
551.48′2—dc21 00-045489

For further information on Blackwell Science, visit our website:
www.blackwell-science.com

Contents

Preface

The editors conceived this volume in response to the growth of available literature on volcaniclastic sedimentation and on lacustrine sedimentation. Although isolated papers have addressed these themes together, this is the first attempt to assemble a group of such contributions, and the first to emphasize how volcanic eruptions can form lakes, influence or control deposition in lakes, and indeed be greatly modified themselves by occurring within lakes. In addition to support from the International Association of Sedimentologists, which provided leadership, editorial assistance, and the interface between volume editors and Blackwell Science, the volume has been sponsored by the Commission on Volcaniclastic Sedimentation (CVS) of the International Association for Volcanology and Chemistry of the Earth's Interior (IAVCEI), through which we invited papers on a broad range of topics in the fields of lacustrine volcaniclastic sedimentation. We were very pleased with the response to our invitation, and, 3 years later, are proud to present the results in this volume.

We are deeply indebted both to the authors who have contributed to the volume and patiently awaited its completion, and to the reviewers who returned manuscripts promptly and whose comments and insight invariably improved the quality of the papers included here. Vern Manville kindly provided a modified version of a diagram from his contribution as a model for the volume's cover art. Finally, we heartily thank Guy Plint, IAS Special Publications editor, who showed phenomenal patience as we chased final authors and reviewers to the close of the process.

The reviewers are:
Beth Palmer, Northfield, MN, USA
David Gaylord, Pulman, WA, USA
Rebecca Dorsey, Eugene, OR, USA
Vern Manville, Taupo, NZ
Sarah Metcalfe, Edinburgh, UK
Larry Middleton, Flagstaff, AZ, USA
Patrick Pringle, Olympia, WA, USA
Stephan Königer, Würzburg, Germany
Nick Foit, Pulman, WA, USA
William Fritz, Atlanta, GA, USA
Gary Smith, Albuquerque, NM, USA
Peter Kokelaar, Liverpool, UK
Reinhard Werner, Kiel, Germany
Jocelyn McPhie, Hobart, Australia
Mike Talbot, Bergen, Norway
R. V. Fisher, Santa Barbara, CA, USA
Grant Heiken, Los Alamos, NM, USA
David Lowe, Hamilton, NZ
Christoph Breitkruz, Potsdam, Germany
Peter Ballance, Auckland, NZ
Richard Hanson, Fort Worth, TX, USA
Chuck Landis, Dunedin, NZ
Thor Thordarson, Perth, Australia
Roger Suthren, Oxford, UK
John Smellie, Cambridge, UK
Dorrik Stowe, Southampton, UK
Bruce Houghton, Wairakei, NZ
Mario Mazzoni (deceased), La Plata, Argentina

Spec. Publs int. Ass. Sediment. (2001) **30**, 1–6

Introduction: styles and significance of lacustrine volcaniclastic sedimentation

J. D. L. WHITE* *and* N. R. RIGGS†

**Geology Department, University of Otago, PO Box 56, Dunedin, New Zealand;*
†Geology Department, Northern Arizona University, Box 4099, Flagstaff, AZ 86011, USA

INTRODUCTION

The focal point of this volume, *Volcaniclastic Sedimentation in Lacustrine Settings*, is the lacustrine depositional record of volcanism. Lakes are a common feature in volcanic terranes, and their deposits are distinctive and useful for interpreting landscape evolution, as well as for trying to gain a systematic understanding of the behaviour of volcanic clasts and dispersal processes. The lacustrine depositional record includes contributions from eruptions taking place far upwind of lakes and their catchment areas, from intra-catchment eruptions that substantially alter lacustrine depositional systems, and from eruptions that take place in, and are partly shaped by, standing water in lakes. Pumiceous lake sediments show a range of features that do not readily fit into facies schemes developed in lakes unaffected by volcanism. Indeed, lacustrine depositional systems offer a natural laboratory for separating the effects of grain size from those of grain mass during deposition and transport of vesicular volcanic fragments.

Lacustrine successions are particularly interesting components of volcanic regions for at least four reasons:
1 the formation of lakes in many cases is a direct or indirect effect of volcanic eruptions, which can result in obstruction of streams, melting of ice, or creation of deep topographic depressions such as calderas and maars;
2 standing water strongly modifies the style of terrestrial eruptions and dispersal of tephra, and can result in pyroclastic deposits unique among terrestrial volcanic sequences;
3 the aqueous sedimentation of volcanic clasts, which vary strongly in density and thus settling behaviour, is of general sedimentological interest and most directly investigated in deposits from standing water;
4 lakes include uniquely low-energy terrestrial sedimentary environments, in which the most detailed and distal records of volcanic fall can be preserved.

VOLCANIC LAKES

Lakes form as a result of volcanic eruptions in a variety of ways. Perhaps the most direct is by the conversion of ice to water by volcanic heat, which forms englacial and subglacial lakes in response to eruptions such as those discussed in this volume by Smellie (pp. 9–35; see also Jones, 1969, 1970; Skilling, 1994; Smellie & Skilling, 1994; Smellie & Hole, 1997). In addition, because glaciers themselves can effectively impound lakes, there may be complex interrelationships among volcanism, development of lakes, and the sedimentary record of this interplay (Werner *et al*., 1996).

Another readily envisioned process by which volcanic eruptions can form lakes is the damming of streams by lava flows. Although this process is surprisingly ineffective in many situations because of the high permeability of jointed and flow-brecciated lavas (Segerstrom, 1950; Young & Jones, 1984; White, 1991; Hamblin, 1994), large lava flows may effectively impound lakes for long periods. This process seems to be particularly effective in silicic volcanic fields (Nairn, 1989; Palmer & Shawkey, 1997). The resulting lakes commonly provide key information for unravelling the eruptive histories of surrounding volcanoes, as shown by the contribution of Palmer & Shawkey in this volume (pp. 179–199).

Streams can also be dammed directly by pyroclastic eruptions, which may disrupt drainage over large areas. Small-volume eruptions may temporarily raise the level of existing lakes (White *et al*., 1997), but large eruptions may result in much larger-scale reorganizations of drainage patterns (Buesch, 1991). As a result of the 1.8 ka eruption of Taupo volcano in New Zealand, ignimbrite not only dammed the outlet to Taupo caldera itself (Wilson & Walker, 1985; White *et al*. this volume, pp. 141–151), but also impounded significant temporary lakes both adjacent to (Smith, 1991) and well beyond the caldera (Manville *et al.*, 1999; Manville, this volume, pp. 109–141). An earlier Taupo Volcanic Zone eruption produced the Rotoiti ignimbrite from Okataina caldera (Nairn, 1989). The ignimbrite dammed the outlet to Lake Rotorua (which occupies a separate caldera), raising the lake level by 90 m and impounding it at that level for some 20 kyr (Kennedy, 1994).

OTHER LAKES IN VOLCANIC ENVIRONMENTS

Not all lakes in volcanic environments owe their origin to volcanic processes, and this is particularly true in active convergent and rifting plate-margin settings, where volcanism and tectonic subsidence are separate but spatially coincident manifestations of lithospheric-scale processes. In their contribution here, Gaylord *et al.* (pp. 199–225) show that volcanogenic lake sedimentation was favoured in the highly extended region represented by the Republic Graben of Washington state because of a combination of rapid basin subsidence, moist climatic conditions, an abundant supply of loose volcanic detritus, and the topographically elevated and isolated nature of the Okanogan Highlands.

LACUSTRINE ERUPTIONS

The first section of the volume, Eruptions and Eruption-formed Lakes, is represented by Smellie's account (pp. 9–35) of intraglacial eruption and deposition of a volcaniclastic suite within the eruption-formed lake. He presents a new model for the ways in which glacial hydrology affects impoundment of intraglacial lakes and the resulting drainage pathways (see also Smellie, 2000a). Smellie's model has wide-ranging implications for analysis of volcaniclastic successions in ice-dammed lakes of any origin.

Once subglacial eruptions have melted enclosing ice to form a body of standing water, eruption processes converge with those typical of other subaqueous, 'Surtseyan', eruptions in which abundant water interacts with erupting magma to produce distinctive cypressoid jets and steam-laden eruption plumes (White & Houghton, 2000). Lakes in which Surtseyan eruptions occur may themselves be a result of volcanic activity, as is the case for the eruption and deposits discussed by Belousov & Belousova (pp. 35–61) that developed in the caldera lake occupying the edifice of Karymskoye volcano. In 1996, seismic activity alerted the Kamchatka Volcanological Observatory to activity at Karymskoye caldera. A helicopter overflight revealed repeated bursts of ash-laden water and laterally expanding steam currents from the previously ice-covered lake. Subsequent investigation showed that the eruption caused tsunami (seiches) to travel across the lake, overflowing into the Karymskaya river to form downstream lahars (debris flows and hyperconcentrated flows) and floods. A small tuff ring was built above the lake surface by the eruption, and depositional features of the ring indicate distribution of ash and other debris, including large blocks of the ice that covered the lake before eruption, by a variety of mechanisms as first the eruption and then the volcano itself shoaled above the lake.

Eruptions initiated beneath non-volcanic lakes share the wide variety of eruption processes, but additionally may provide unique insights into the history of the enclosing lake itself. Pahvant Butte volcano erupted into Lake Bonneville near its highstand level (Gilbert, 1890; Oviatt & Nash, 1989; White, this volume, pp. 61–83), and ash from the eruption provides a time plane and a marker of shoreline position unparalleled in the Bonneville basin. Reworked ash on the volcano itself is a sensitive water-level indicator, and provides critical support for a late lake-level oscillation in Lake Bonneville just before its catastrophic breakout into the Snake River catchment (Gilbert, 1890; Spencer *et al.*, 1984; Sack, 1989). Relatively deep standing water (≈ 85 m) at the eruption site produced a distinctive suite of eruption-fed clastic deposits that built up from the floor of the lake to its surface. It is inferred that material erupted from the subaqueous vent was dispersed upward into the water column, then entrained into dilute aqueous density currents, turbidity currents, which dispersed the debris laterally to form extensive subhorizontal beds of well-sorted ash. The source characteristics of such density currents are unique in their combination of intermittency and sediment dispersion.

The contribution by Cas *et al.* (pp. 83–109) addresses deposits formed in a pre-existing lake in which predominantly rhyolitic magma erupted both subaqueously within the lake basin and subaerially, affecting the catchments beyond the lake. Eruptions included both explosive and non-explosive phases (Cas *et al.*, 1990), with the latter dominant in the lake. Volcaniclastic deposits formed from direct volcanic fallout, from eruption-related sediment gravity flows, and from sediment gravity flows occurring between eruptions that were sourced from within the lake.

These are interbedded with lacustrine suspension and density-current deposits composed of non-volcanic sediment and of remobilized volcanic debris, both carried by streams to the lake from surrounding catchment areas.

ERUPTION-IMPOUNDED LAKES

Uniquely among this group of papers, Manville describes in this volume (pp. 109–141) the deposits of a lake that formed as a direct result of a large eruption from a nearby volcano. Emplacement of the Taupo ignimbrite profoundly disrupted the Waikato River catchment area near the volcano to form Lake Reporoa, which covered a large area to depths of tens of metres yet is inferred to have formed, filled, overflowed, and been drained over the course of less than a decade. During its brief existence, a full suite of very strongly pumiceous deposits, including deltaic successions and basin-centre turbidites, was formed.

Enlargement of Lake Taupo, albeit following its almost complete emptying during the course of a large-scale intralacustrine eruption (Wilson & Walker, 1985), resulted in deposition of a transgressive accumulation of highly pumiceous ignimbrite-derived pyroclastic debris at shoreline to water depths of tens of metres in a zone surrounding the present lake. White *et al.* (this volume, pp. 141–151) investigate distinctive features of these deposits to develop a general assessment of lacustrine pumice-deposition processes. In their contribution, Riggs *et al.* (pp. 151–179) detail specific depositional environments developed around the raised lake and interpret the different lithofacies developed in terms of a balance among the rate of accommodation development, wave and current energy, and sediment influx; draining of the lake from its post-eruptive highstand to the present level is also briefly addressed, and Manville *et al.* (1999) have modelled the resulting outbreak flood.

SEDIMENTATION AND RESEDIMENTATION OF PYROCLASTS

As volcaniclastic deposits became a subject of specific study, investigators having backgrounds in stratigraphy, sedimentology, and volcanology converged upon them. This fertile mix of expertise has driven rapid advances in understanding of such deposits, but has also resulted in inconsistent and overlapping sets of terminology applied by different workers. Our own preference is for terminology following Fisher & Schmincke (1984), which is based upon grain origin. Hence a pyroclast is a fragment formed during an eruption, and a rock made of such fragments is named as a pyroclastic rock (e.g. 'tuff') without regard to whether the pyroclasts form a 'primary' fall deposit or have been 'redeposited' by aqueous or aeolian processes. A major alternative point of view is that pyroclasts worked by water are best considered sedimentary particles, and their products named as sedimentary rocks (Cas & Wright, 1987; McPhie *et al.*, 1993). In this volume each contributor has made clear how terms are used.

Section 2, Resedimentation in Lakes, includes papers examining the dispersal and sedimentation within lakes of volcaniclastic debris from a variety of largely extralacustrine eruptive sites. In addition to rhyolitic debris supplied by intralacustrine eruptions, Cas *et al.* (pp. 83–109) describe here how deposits of the Bunga Beds also contain material formed by eruptions alongside the lake and then carried into the lake via fluvio-deltaic systems. Closely linked with eruptions on land are 'continuous-feed' turbidity currents, which are postulated to have produced some of the turbidites in response to hyperpycnal outflows that may represent the basinal extension of lahars.

Climate and tectonism are considered the primary controls on sedimentation globally (e.g. Bridge & Leeder, 1979; DeCelles *et al.*, 1991; Bettis & Autin, 1997). The contribution by Palmer & Shawkey (pp. 179–199) develops this theme by describing lacustrine products of volcanism in the context of these broader controls. Eocene deposits that accumulated in a small intermontane basin in central Idaho, USA, record with remarkable sensitivity the interplay between tectonism, climate, and volcanic activity. Interstratified lacustrine and fluvial deposits document changes in sediment supply, base level, and discharge. Gaylord *et al.* (pp. 199–225), in a complementary but unrelated paper, document the facies array that accumulated in a rapidly subsiding graben that formed during Eocene extension along the US Cordillera. As topography was lowered while metamorphic core complexes rose as a result of extension in western North America, basins that developed over the complexes had remarkable preservation potential. Because of the close proximity of active volcanoes to the lacustrine depositional sites, Gaylord *et al.* were able to document the specific contribution of hydrothermal alteration to processes of disaggregation and incorporation of a variety of products to the sedimentary systems.

LAKES AS TEPHRASTRATIGRAPHIC REPOSITORIES

Section 3 concludes the volume with papers focused on the role of lakes as distal repositories of volcanic ash, their use in reconstructing prehistoric eruptions and environmental conditions, and the techniques used to best utilize this unique, but not always uniquely interpretable, data set. Widespread ash layers useful for such work commonly arise from extrabasinal eruptions, from which ash is spread over downwind areas that may extend across hundreds of thousands of square kilometres (Walker, 1973). Such layers may extend across both terrestrial and marine environments, and offer the most precise available means of correlating terrestrial and marine biostratigraphy (e.g. Carter *et al.*, 1995).

In contrast to this assessment of how tephra horizons can be used to provide widespread stratigraphic correlations, the contribution of Hardardóttir *et al.* (pp. 225–247) details how normal biogenic and biochemical lacustrine processes are interrupted by fallout accumulation. Throughout Iceland, lakes provide a detailed record of volcanism, but from a detailed study of Lake Hestvatn, these researchers show how complex environmental changes normally reflected in diatom productivity and pollen populations must be interpreted through the filter of tephra depositional processes.

A similar technique is applied by Caballero *et al.* (pp. 247–263) to understanding the evolution of Nevado de Toluca over the past 40 kyr. In this case, longer-term activity from Toluca had more long-lasting effects on the size and shape of the lake than on the biological processes taking place within it. Not surprisingly, these researchers are best able to document environmental changes, in part brought on by volcanic activity, during non-volcanic periods.

Erosion and sediment redistribution are ubiquitous in terrestrial settings, and lacustrine deposits containing age-diagnostic material are exceptionally useful tools for chronostratigraphic correlation. Pyroclastic deposits are not only commonly datable by isotopic means, but also form lithostratigraphic units that may extend across widely varying depositional settings (Fisher & Schmincke, 1984). Königer & Stollhofen show in this volume (pp. 263–285) how pyroclastic material, preserved as both ‘primary’ waterlain fall deposits in lakes and ‘reworked’ ash in associated fluvial settings, can be used to develop a chronostratigraphic and lithostratigraphic framework for otherwise disjunct terrestrial units of Permo-Carboniferous age in the Saar–Nahe basin. Ash layers formed during lacustrine transgressions, and on footwall blocks of the rift basin, are commonly primary waterlain fall deposits, whereas equivalent tephra deposited in other surroundings exists only in reworked form, in many instances mixed with non-volcanic sediment.

Information derived from study of tephra produced by a single eruption and deposited in a landslide-dammed lake downwind of Crater Lake, Oregon, has been used by Riedel *et al.* (pp. 285–299) in their contribution to infer post-eruption lacustrine accumulation rates in excess of 15 m yr^{-1}. They also show that a regionally 2-cm-thick layer of tephra, preserved as a waterlain ash-fall bed in the lake, was eroded from catchment hillslopes and redeposited to form > 10 m of lacustrine suspension deposits and turbidites.

CONCLUSION

An impressive range of geological information, deposit types, temporal duration and geographical extent, and approaches to analysis is manifest in the studies presented here. The key feature of lacustrine volcaniclastic sedimentation is that it responds, often strongly, to an allogenic forcing mechanism that does not affect lakes uninfluenced by volcanoes. Studies of the resulting successions benefit from a multidisciplinary approach, and in turn illuminate both the response of the forced lacustrine setting and aspects of the volcanic forcing mechanisms themselves.

REFERENCES

Belousov, A. & Belousova, M. (2000) Eruptive process, effects and deposits of the 1996 and ancient basaltic phreatomagmatic eruptions in Karymskoye lake, Kamchatka, Russia. In: *Volcaniclastic Sedimentation in Lacustrine Settings* (Eds White, J.D.L. & Riggs, N.R.), Spec. Publs int. Assoc. Sediment., No. 30, pp. 35–61. Blackwell Science, Oxford.

Bettis, E.A., III & Autin, J.W. (1997) Complex response of a midcontinent North American drainage system to late Wisconsinan sedimentation. *J. sediment. Res.*, **67**, 740–748.

Bridge, J.S. & Leeder, M.R. (1979) A simulation model of alluvial stratigraphy. *Sedimentology*, **26**, 617–644.

Buesch, D.C. (1991) Changes in depositional environments resulting from emplacement of a large-volume ignimbrite. In: *Sedimentation in Volcanic Settings* (Eds Fisher, R.V. & Smith, G.A.), Spec. Publ. Soc. econ. Paleont. Miner., Tulsa, **45**, 139–153.

Caballero, M., Macías, J.L., Lozano-García, S., Urrutia-Fucugauchi, J. & Casteñada-Bernal, R. (2000) Late

Pleistocene–Holocene volcanic stratigraphy and palaeoenvironments of the upper Lerma basin, Mexico. In: *Volcaniclastic Sedimentation in Lacustrine Settings* (Eds White, J.D.L. & Riggs, N.R.), Spec. Publs int. Assoc. Sediment., No. 30, pp. 247–263. Blackwell Science, Oxford.

Carter, L., Nelson, C.S., Neil, H.L. & Froggatt, P.C. (1995) Correlation, dispersal, and preservation of the Kawakawa Tephra and other late Quaternary tephra layers in the Southwest Pacific Ocean. *N.Z. J. Geol. Geophys.*, **38**, 29–46.

Cas, R.A.F. & Wright, J.V. (1987) *Volcanic Successions, Modern and Ancient*. Allen and Unwin, London.

Cas, R.A.F., Allen, R.L., Bull, S.W., Clifford, B.A. & Wright, J.V. (1990) Subaqueous, rhyolitic dome-top tuff cones: a model based on the Devonian Bunga Beds, southeastern Australia and a modern analogue. *Bull. Volcanol.*, **52**, 159–174.

Cas, R.A.F., Edgar, C., Allen, R.L., Bull, S., Clifford, B.A., Giordano, G. & Wright, J.V. (2000) Influence of magmatism and tectonics on sedimentation in an extensional lake basin: the Upper Devonian Bunga Beds, Boyd Volcanic Complex, southeastern Australia. In: *Volcaniclastic Sedimentation in Lacustrine Settings* (Eds White, J.D.L. & Riggs, N.R.), Spec. Publs int. Assoc. Sediment., No. 30, pp. 83–109. Blackwell Science, Oxford.

Decelles, P.G., Gray, M.B., Ridgeway, K.D., *et al.* (1991) Controls on synorogenic alluvial-fan architecture, Beartooth Conglomerate (Paleocene), Wyoming and Montana. *Sedimentology*, **38**, 567–590.

Fisher, R.V. & Schmincke, H.-U. (1984) *Pyroclastic Rocks*. Springer, Berlin.

Gaylord, D.R., Price, S.M. & Suydam, J.D. (2000) Volcanic and hydrothermal influences on middle Eocene lacustrine sedimentary deposits, Republic basin, northern Washington, USA. In: *Volcaniclastic Sedimentation in Lacustrine Settings* (Eds White, J.D.L. & Riggs, N.R.), Spec. Publs int. Assoc. Sediment., No. 30, pp. 199–225. Blackwell Science, Oxford.

Gilbert, G.K. (1890) *Lake Bonneville.* Monogr. US geol. Surv., Denver, CO, **1**.

Hamblin, W.K. (1994) *Late Cenozoic Lava Dams in the Western Grand Canyon*. Geol. Soc. Am. Mem., Boulder, CO, **183**.

Hardardóttir, J., Geirsdóttir, Á. & Thórdarson, T. (2000) Tephra layers in a sediment core from Lake Hestvatn, southern Iceland: implications for evaluating sedimentation processes and environmental impacts on a lacustrine system caused by tephra fall deposits in the surrounding watershed. In: *Volcaniclastic Sedimentation in Lacustrine Settings* (Eds White, J.D.L. & Riggs, N.R.), Spec. Publs int. Assoc. Sediment., No. 30, pp. 225–247. Blackwell Science, Oxford.

Jones, J.G. (1969) Intraglacial volcanoes of the Laugarvatn region, southwest Iceland. *J. geol. Soc. London*, **124**, 197–211.

Jones, J.G. (1970) Intraglacial volcanoes of the Laugarvatn region, southwest Iceland II. *J. Geol.*, **78**, 127–147.

Kennedy, N. (1994) New Zealand tephro-chronology as a tool in geomorphic history of the *c*. 140 ka Mamaku Ignimbrite Plateau and in relating oxygen isotope stages. *Geomorphology*, **9**, 97–115.

Königer, S. & Stollhofen, H. (2000) Environmetal and tectonic controls on preservation potential of distal fallout ashes in fluvio-lacustrine settings: the Carboniferous–Permian Saar–Nahe Basin, SW Germany. In: *Volcaniclastic Sedimentation in Lacustrine Settings* (Eds White, J.D.L. & Riggs, N.R.), Spec. Publs int. Assoc. Sediment., No. 30, pp. 263–285. Blackwell Science, Oxford.

Manville, V. (2000) Sedimentology and history of Lake Reporoa: an ephemeral supra-ignimbrite lake, Taupo Volcanic Zone, New Zealand. In: *Volcaniclastic Sedimentation in Lacustrine Settings* (Eds White, J.D.L. & Riggs, N.R.), Spec. Publs int. Assoc. Sediment., No. 30, pp. 109–141. Blackwell Science, Oxford.

Manville, V., White, J.D.L., Houghton, B.F. & Wilson, C.J.N. (1999) Paleohydrology and sedimentology of a post-1.8 ka breakout flood from intracaldera Lake Taupo, North Island, New Zealand. *Geol. Soc. Am. Bull.*, **111**, 1435–1447.

McPhie, J., Doyle, M. & Allen, R. (1993) *Volcanic Textures: a Guide to the Interpretation of Textures in Volcanic Rocks*. CODES Key Centre, University of Tasmania, Hobart.

Nairn, I.A. (1989) Sheet V16AC Tarawera; Geological Map of New Zealand, 1 : 50 000. Department of Scientific and Industrial Research, Wellington.

Oviatt, C.G. & Nash, W.P. (1989) Late Pleistocene basaltic ash and volcanic eruptions in the Bonneville basin, Utah. *Geol. Soc. Am. Bull.*, **101**, 292–303.

Palmer, B.A. & Shawkey, E.P. (1997) Lacustrine sedimentation processes and patterns during effusive and explosive volcanism, Challis volcanic field, Idaho. *J. sediment. Res.*, **67**, 154–167.

Palmer, B.A. & Shawkey, E.P. (2000) Lacustrine–fluvial transitions in a small intermontane valley, Eocene Challis volcanic field, Idaho. In: *Volcaniclastic Sedimentation in Lacustrine Settings* (Eds White, J.D.L. & Riggs, N.R.), Spec. Publs int. Assoc. Sediment., No. 30, pp. 99–118. Blackwell Science, Oxford.

Riedel, J.L., Pringle, P.T. & Schuster, R.L. (2000) Deposition of Mount Mazama tephra in a landslide-dammed lake on the upper Skagit River, Washington, USA. In: *Volcaniclastic Sedimentation in Lacustrine Settings* (Eds White, J.D.L. & Riggs, N.R.), Spec. Publs int. Assoc. Sediment., No. 30, pp. 285–299. Blackwell Science, Oxford.

Riggs, N.R., Ort, M.H., White, J.D.L., Wilson, C.J.N., Houghton, B.F. & Clarkson, R. (2000) Post-1.8-ka marginal sedimentation in Lake Taupo, New Zealand: effects of wave energy and sediment supply in a rapidly rising lake. In: *Volcaniclastic Sedimentation in Lacustrine Settings* (Eds White, J.D.L. & Riggs, N.R.), Spec. Publs int. Assoc. Sediment., No. 30, pp. 151–179. Blackwell Science, Oxford.

Sack, D. (1989) Reconstructing the chronology of Lake Bonneville: an historical review. In: *History of Geomorphology: from Hutton to Hack* (Ed. Tinkler, K.J.), Binghamton Symposia in Geomorphology, International Series, Vol. 19, pp. 223–256. Unwin Hyman, Boston, MA.

Segerstrom, K. (1950) Erosion studies at Paricutin volcano, State of Michoacan, Mexico. *US geol. Surv. Bull.*, **956-A**, 1–164.

Skilling, I.P. (1994) Evolution of an englacial volcano: Brown Bluff, Antarctica. *Bull. Volcanol.*, **56**, 573–591.

Smellie, J.L. (2000a) Subglacial eruptions. In: *Encyclopedia of Volcanoes* (Eds Sigurdsson, H., Houghton, B.F., McNutt, S.R., Rymer, H. & Stix, J.), pp. 403–418. Academic Press, San Diego, CA.

Smellie, J.L. (2000b) Lithofacies architecture and construction of volcanoes erupted in englacial lakes: Icefall Nunatak, Mount Murphy, eastern Marie Byrd Land, Antarctica. In: *Volcaniclastic Sedimentation in Lacustrine Settings* (Eds White, J.D.L. & Riggs, N.R.), Spec. Publs int. Assoc. Sediment., No. 30, pp. 9–35. Blackwell Science, Oxford.

Smellie, J.L. & Hole, M.J. (1997) Products and processes in Pliocene–Recent, subaqueous to emergent volcanism in the Antarctic Peninsula: examples of englacial Surtseyan volcano construction. *Bull. Volcanol.*, **58**, 628–646.

Smellie, J.L. & Skilling, I.P. (1994) Products of subglacial volcanic eruptions under different ice thicknesses: two examples from Antarctica. *Sediment. Geol.*, **91**, 115–129.

Smith, R.C.M. (1991) Landscape response to a major ignimbrite eruption, Taupo Volcanic Center, New Zealand. In: *Sedimentation in Volcanic Settings* (Eds Fisher, R.V. & Smith, G.A.), Spec. Publ. Soc. econ. Paleont. Miner., Tulsa, **45**, 123–137.

Spencer, R.J., Baedecker, M.J., Eugster, H.P., *et al.* (1984) Great Salt Lake, and precursors, Utah: the last 30,000 years. *Contrib. Mineral. Petrol.*, **86**, 321–334.

Walker, G.P.L. (1973) Explosive volcanic eruptions—a new classification scheme 11. *Geol. Rundsch.*, **62**, 431–446.

Werner, R., Schmincke, H.-U. & Sigvaldason, G. (1996) A new model for the evolution of table mountains: volcanological and petrological evidence from the Herdubreid and Herdubreidarotgl volcanoes (Iceland). *Geol. Rundsch.*, **85**, 390–397.

White, J.D.L. (1991) The depositional record of small, monogenetic volcanoes within terrestrial basins. In: *Sedimentation in Volcanic Settings* (Eds Fisher, R.V. & Smith, G.A.), Spec. Publ. Soc. econ. Paleont. Miner., Tulsa, **45**, 155–171.

White, J.D.L. (2000) Eruption and reshaping of Pahvant Butte volcano in Pleistocene Lake Bonneville. In: *Volcaniclastic Sedimentation in Lacustrine Settings* (Eds White, J.D.L. & Riggs, N.R.), Spec. Publs int. Assoc. Sediment., No. 30, pp. 61–83. Blackwell Science, Oxford.

White, J.D.L. & Houghton, B.F. (2000) Surtseyan and related eruptions. In: *Encyclopedia of Volcanoes* (Eds Sigurdsson, H., Houghton, B., McNutt, S., Rymer, H. & Stix, J.), pp. 495–512. Academic Press, San Diego, CA.

White, J.D.L., Houghton, B.F., Hodgson, K.A. & Wilson, C.J.N. (1997) Delayed sedimentary response to the 1886 a.d. eruption of Tarawera, New Zealand. *Geology*, **25**, 459–462.

White, J.D.L., Manville, V., Wilson, C.J.N., Houghton, B.F., Riggs, N.R. & Ort, M. (2000) Settling and deposition of AD 181 Taupo pumice in lacustrine and associated environments. In: *Volcaniclastic Sedimentation in Lacustrine Settings* (Eds White, J.D.L. & Riggs, N.R.), Spec. Publs int. Assoc. Sediment., No. 30, pp. 141–151. Blackwell Science, Oxford.

Wilson, C.J.N. & Walker, G.P.L. (1985) The Taupo eruption, New Zealand I. General aspects. *Phil. Trans. R. Soc. London, Ser. A*, **314**, 199–228.

Young, H.W. & Jones, M.L. (1984) *Hydrologic, demographic, and land-use data for the Snake River Plain, southeastern Idaho.* US geol. Surv. Water-Resour. Invest. Rep. **84-4001**.

Eruptions and eruption-formed lakes

Spec. Publs int. Ass. Sediment. (2001) **30**, 9–34

Lithofacies architecture and construction of volcanoes erupted in englacial lakes: Icefall Nunatak, Mount Murphy, eastern Marie Byrd Land, Antarctica

J. L. SMELLIE

British Antarctic Survey, High Cross, Madingley Road, Cambridge CB3 0ET, UK

ABSTRACT

Mount Murphy is a large Miocene shield volcano flanked by several small basaltic satellite centres that were erupted beneath a thick (> 200 m) ice sheet. Three empirical models illustrating the hydraulic evolution of glacio-volcanic systems are deduced from glacier physics, with distinctly different implications in each case for the resultant lithofacies architecture. Glacier hydraulic considerations and facies analysis are used to describe the evolution of one of the satellite centres (Icefall Nunatak). The nunatak was constructed from several vents during three main stages. Each stage demonstrates different aspects of englacial volcano construction, mainly in a flooded vault or lacustrine setting. An initial mainly effusive phase was dominated by lava and cogenetic joint-block breccia, and eruption was probably confined mainly within an englacial vault or lake (stage I). Renewed activity, at a different vent and beneath a re-established ice sheet (stage II), began with coarse sediments flushed away subglacially. A subaqueous tuff cone was then constructed in an englacial lake, from explosively erupted coarse glassy tephra probably produced mainly during sustained eruptions and distributed by high-density turbidity currents. Fine detritus is common only in the basal tuff cone unit, possibly as a result of lower, denser (largely subaqueous?) eruption columns. A spectacular slope failure is represented by numerous large blocks, which were displaced to low elevations on extensively fractured tuff cone flanks, and the failure event may have initiated zones of high pore-water discharge. Stage II culminated with two phases of lava delta progradation, indicating that the volcanic edifice ultimately penetrated the entire ice-sheet thickness and that the vent became emergent. Stage III commenced with lava effusion, probably through a thin re-formed cover of permeable snow and firn. A small cinder cone was also constructed and was partially palagonitized because of its structural position on top of a water-saturated volcanic pile and likely presence of vent intrusions driving hydrothermal circulation.

INTRODUCTION

Volcanoes that form as a result of explosive eruptions in lakes represent active high-energy systems with rapid episodic (linked to eruptions) input of coarse-grained tephra to the high-relief steep submerged volcano flanks (e.g. Sohn, 1995; White, 1996; Smellie & Hole, 1997). Erosional modification in exposed settings can complicate the depositional record of lacustrine volcanoes (e.g. White, 1996) and most lakes in volcanic settings are depocentres for sedimentary (volcanogenic epiclastic) detritus in addition to tephra. Identification of primary volcanic processes depends on the successful screening out of non-volcanic influences. Distinguishing purely epiclastic sedimentation (Cas & Wright, 1987) of a volcanic provenance from redeposited syneruptive tephra in lake successions can be difficult, yet it is important if we are to interpret the palaeoenvironmental and eruptive records preserved in ancient lake successions.

Volcanoes erupted in lakes are also invaluable sources of information on hydrovolcanism. However, uplifted exposed sections are comparatively uncommon and few studies give details of the lithofacies present and their relationships (e.g. White, 1996). By contrast, subglacially erupted volcanoes, which are constructed largely within englacial lakes, are common and often well exposed, particularly in Antarctica (e.g. Hamilton, 1972; LeMasurier, 1972; Wörner & Viereck, 1987; Smellie *et al*., 1988, 1993a,b; LeMasurier *et al*., 1994; Skilling, 1994; Smellie & Skilling, 1994; Smellie & Hole, 1997). Unlike other lacustrine systems, which receive variable influxes of sedimentary detritus from the surrounding (non-glacial) terrain, englacial lakes formed around active volcanoes are essentially isolated physically by the surrounding ice from most external (non-volcanic) sediment sources. Englacial lakes are also very protected settings and erosional modification caused by waves or strong currents is typically minor.

This paper explores the hydrodynamic background of eruptions within glaciers and the results are illustrated by examining the lithofacies architecture and construction of a small, very well exposed polygenetic volcanic centre of late Miocene age at Icefall Nunatak, near Mount Murphy, eastern Marie Byrd Land. The centre displays the varied eruptive and depositional processes characteristic of basaltic englacial lacustrine volcanoes, which are also known as table-mountain or tuya volcanoes (e.g. Jones, 1969, 1970; Allen *et al*., 1982). In these volcanoes, subaqueous (pillow lava, tuff cone) lithofacies are overlain by hyaloclastite deltas as a consequence of shoaling and emergence of the vents. Activity is Surtseyan during the emergent period, when explosive eruptions take place in a flooded vent (Kokelaar, 1983) and, in many aspects of their construction and eruptive characteristics, these englacial volcanoes are generally similar to Surtseyan volcanoes in marine situations (Jones, 1966; Smellie & Hole, 1997). The study also illustrates the fundamental controls of glacier hydrology on the sequence of events and depositional record of hydrovolcanic eruptions in englacial environments.

GEOLOGICAL BACKGROUND

Volcanism in Marie Byrd Land extends back to Late Oligocene times (28 Ma), at least, and coincided with the development of a widespread Antarctic ice sheet (LeMasurier, 1972; LeMasurier & Rex, 1982). The volcanism is alkaline, related to the impingement of a major mantle plume beneath the stationary Marie Byrd Land crustal microplate and coincident development of a basin-and-range-like extensional province in the Ross embayment (the West Antarctic rift system) during Cenozoic times (LeMasurier & Rex, 1989; Behrendt *et al*., 1991; Hole & LeMasurier, 1994). The volcanic province contains at least 18 major phonolite and trachyte shield or stratovolcanoes with large (up to 10 km diameter) summit calderas, and numerous smaller basaltic centres (LeMasurier, 1990; Hole & LeMasurier, 1994; Panter *et al*., 1994, 1997). The proportion of pyroclastic deposits is generally small in most of the stratovolcanoes and comprises numerous scoria cones and very rare Plinian fall and pyroclastic flow deposits (Panter *et al*., 1994; Wilch *et al*., 1999). Conversely, hydroclastic deposits (tephra and hyaloclastite autobreccias) are locally abundant and are particularly common in the basal sections of some volcanoes ('basal sequence' of LeMasurier, 1972; see also LeMasurier & Rex, 1982; LeMasurier, 1990).

Mount Murphy is situated in eastern Marie Byrd Land (Fig. 1). It is a large dissected volcano dominated by a basanite to peralkaline trachyte shield ≈ 25 km in diameter with a postulated small (4–5 km diameter) ice-filled summit caldera breached on the south side (LeMasurier, 1990; Figs 1 & 2). The shield was constructed relatively quickly in late Miocene times (mainly 8–9 Ma) on a lithologically varied basement composed of early Cenozoic and older granitoid and gabbroid plutons, alkaline dykes, gneiss and turbidite sedimentary strata. The surface of the local basement is a prominent feature forming a conspicuous massif rising from < 400 m to > 2000 m above sea-level (a.s.l.; Fig. 1). Younger activity was entirely basaltic (basanite, alkali basalt and hawaiite) and comprised several small satellite centres erupted in latest Miocene times (6–7 Ma) and now highly dissected, and a few small Pliocene to Recent tuff and cinder cones.

The Mount Murphy volcano has had a complicated history of interactions between magma and former ice (Smellie *et al*., 1993b; LeMasurier *et al*., 1994). The lower part of the shield succession (between 400 and 1500 m a.s.l.) is dominated by basaltic lavas and hydroclastic deposits formed both explosively and by autoclastic brecciation. It includes interbedded thin diamictites (tillites) resting on multiple glacially eroded striated surfaces and an association with ice coeval with eruptions is undoubted. These initial eruptions were of subglacial 'sheet-flow' type, comparable in many respects with those described by Walker & Blake (1966) and Smellie *et al*. (1993a), and formed when the slopes of the volcano were mantled by relatively thin 'ice' (probably mainly firn and/or snow < 100 m thick). The overlying succession consists almost solely of subaerially erupted basaltic and trachytic lavas. After construction of the shield, several latest Miocene satellite centres (including Icefall Nunatak) were constructed. Unlike the shield, they display multiple subaqueous to subaerial transitions similar to passage zones in seamounts and in table-mountain volcanoes erupted beneath thick ice sheets (greater than ≈ 200 m

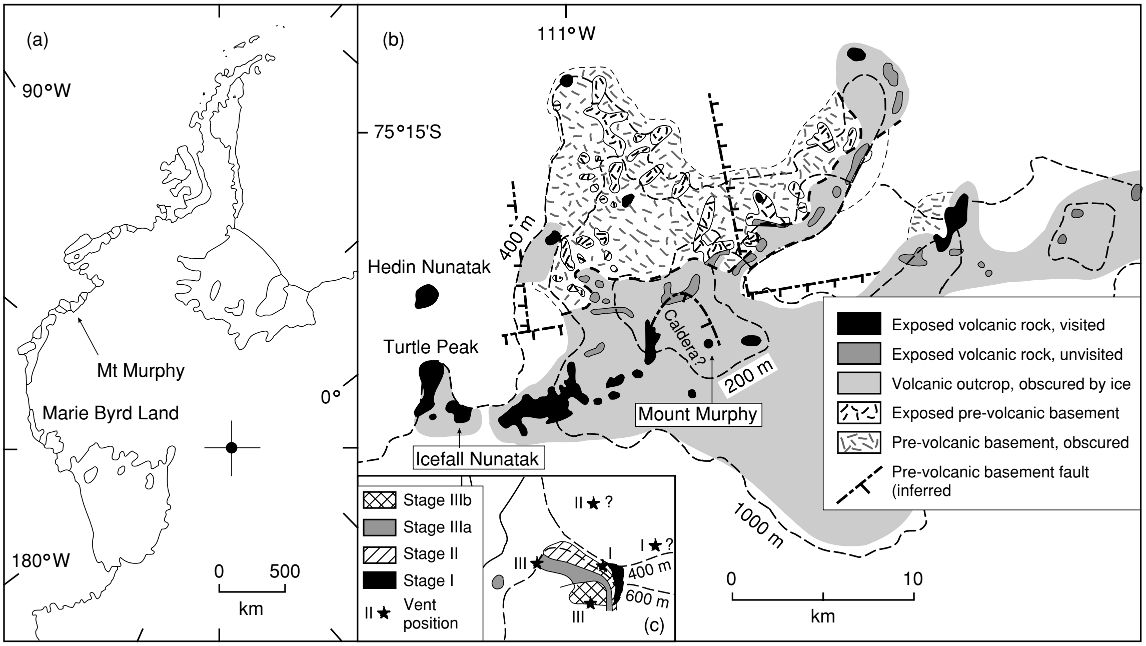

Fig. 1. (a) Sketch map showing location of the study area (Mount Murphy) in Marie Byrd Land. (b) Sketch map showing location of Icefall Nunatak and simplified geology of Mount Murphy. (c) Geological sketch map of Icefall Nunatak, with the stratigraphy separated into the three evolutionary stages described in this paper. The known and inferred locations of vents responsible for constructing the polygenetic volcano at Icefall Nunatak are also shown in (c).

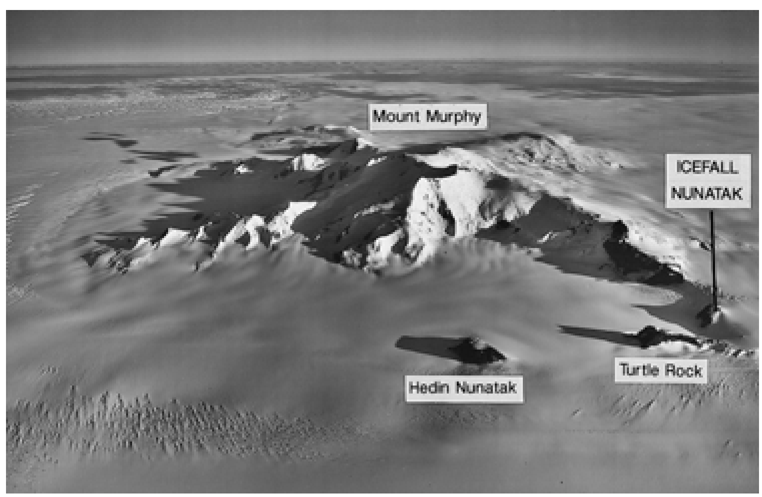

Fig. 2. Aerial view of Mount Murphy and surrounding nunataks, looking east-south-east. Mount Murphy is a large glacially dissected stratovolcano constructed on a north-sloping fault-block massif of pre-Cenozoic 'basement' rocks (Fig. 1), whereas Hedin Nunatak, Turtle Rock and Icefall Nunatak are small basaltic satellite centres. The latter were erupted beneath a gently north-sloping late Miocene ice sheet, which, at times, was perhaps 100 m thicker than that present on the south side of Icefall Nunatak today, and it would also have completely covered Hedin Nunatak and Turtle Rock. US Navy photograph TMA1719 frame F31-132.

thick; see below and Jones, 1966, 1969). From *c.* 3.5 Ma onward, only cinder cones and rare tuff cones were formed. These young pyroclastic cones lack evidence for significant interaction with ice.

Icefall Nunatak is one of the satellite centres situated on the west side of Mount Murphy. It is well exposed in an east–west-trending cliff ≈ 0.9 km long and 200 m high (Fig. 3). The base is concealed by ice and the top corresponds to the present-day erosion surface. The volcano was erupted subglacially in late Miocene times (*c.* 6.5 Ma) and contains evidence for eruptions from several vents.

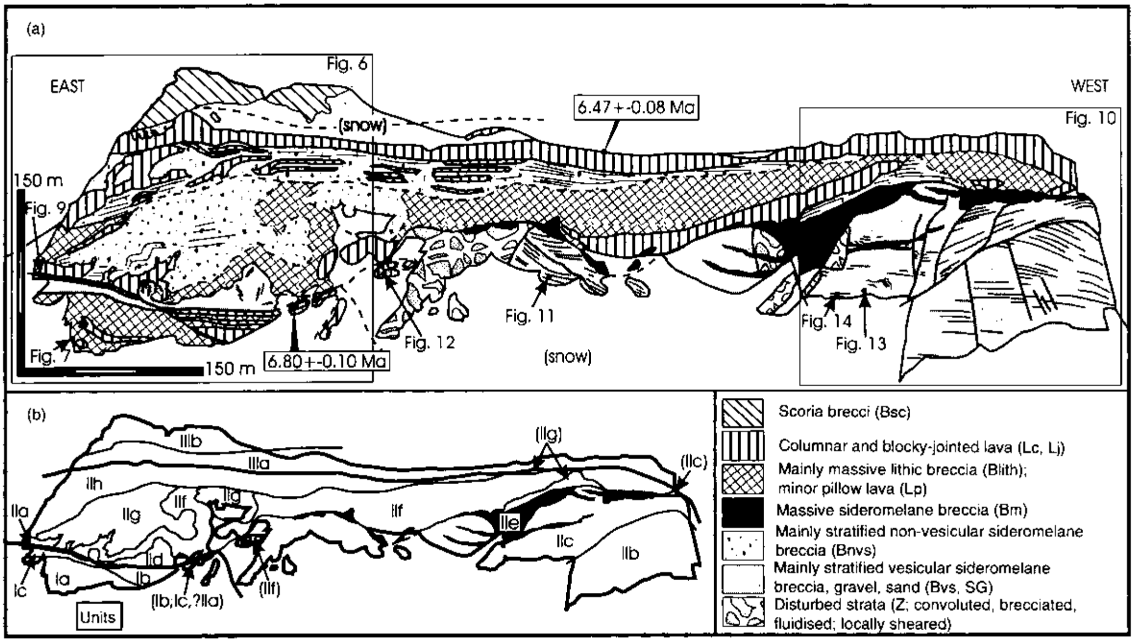

Fig. 3. Sketch of Icefall Nunatak, looking south, showing (a) the distribution of the major lithofacies, and (b) the lithofacies architecture (units described in the text). The locations and ages of dated samples are also shown, and the locations of features illustrated as photographic figures in this paper.

EVIDENCE FOR A GLACIAL SETTING FOR ERUPTIONS AT ICEFALL NUNATAK

Icefall Nunatak is dominated by subaqueous volcaniclastic and volcanic lithofacies, whose characteristics (see below) indicate that they accumulated in ponded water, in either a lacustrine or a marine setting. Apart from one thin unit, the sequences lack the abundant tractional structures indicative of a shallow-marine environment, consistent with a very protected location (lake or deep water (below wave base)). The deposits are also entirely devoid of fossils. Because even rapidly constructed submarine edifices are colonized by marine vegetation and benthic fauna soon after eruption or contain infiltrated marine pelagic fossils (Surtsey Research Society, 1970; Kokelaar & Durant, 1983; Smellie *et al.*, 1998), a non-marine (lacustrine) environment is more likely. However, there is no known Miocene palaeotopography that could have acted as a barrier and confined a former non-glacial lake, suggesting that ice may have acted as a barrier and the lake may have been glacially confined.

That a glacial setting is likely is also suggested by the history of Cenozoic glaciation in the region, which probably commenced at *c.* 40 Ma in East Antarctica and became widespread throughout West Antarctica from early mid-Miocene time (e.g. Cooper *et al.*, 1991). This is supported by field evidence from the Mount Murphy shield succession, which was erupted in association with thin 'ice' in late Miocene times (8–9 Ma) (see above). Moreover, the volcanic sequence at Hedin Nunatak, one of the western satellite centres situated just 6 km north-north-west of Icefall Nunatak (Figs 1 & 2), is composed of multiple superimposed hyaloclastite deltas with passage zones that vary in elevation within a vertical interval of 100 m. The passage zones reflect former water levels coeval with eruptions (Jones, 1969). The entire Hedin Nunatak sequence formed rapidly between 6.50 ± 0.06 and 6.20 ± 0.12 Ma. There is insufficient time for the variations in water level to be an effect of a fluctuating eustatic sea level. This suggests that eruptions were in an englacial lake(s), in association with a relatively thick ice sheet (> 200 m; see Smellie & Skilling, 1994; and see below). The passage zones broadly reflect elevations of the ice-sheet surface (Smellie *et al.*, 1993a). Icefall Nunatak erupted between 6.80 ± 0.10 and 6.47 ± 0.08 Ma (Fig. 3). It therefore erupted between the periods of ice cover represented by the Mount Murphy shield succession and that at Hedin Nunatak. It also overlaps in age with Hedin Nunatak. A glacial eruptive environment and relatively thick ice cover are thus likely for the Icefall Nunatak sequence. From passage zone elevations at Icefall and Hedin nunataks, water levels at the two localities were up to 200 m different at essentially coincident times (≈ 6.5 Ma). These differences are attributed here to eruptions beneath a gently north-sloping ice sheet rather than reflecting a common sea level followed by faulting between the two localities, although the exposures are discontinuous and effects of local tectonism cannot be wholly excluded.

In summary, these various lines of evidence, none of which is independently conclusive, suggest strongly that Icefall Nunatak was formed in a glacial environment, in association with coeval ice. Within that setting, important features, such as the presence of passage zones and volcaniclastic lithofacies deposited in ponded water, are only consistent with eruptions beneath a relatively thick ice sheet and accumulation within an englacial vault or lake.

EFFECTS OF GLACIER PHYSICS ON SUBGLACIAL ERUPTIONS

Eruptions are subglacial when the vents are situated beneath a glacier, whereas the resulting volcanoes are constructed englacially (i.e. surrounded by a glacier; terminology similar to that used by Wright, 1980). Because of rapid (volcano-induced) melting, the volcanoes are typically surrounded by meltwater confined in an englacial vault or lake (Jones, 1969, 1970; Smellie & Skilling, 1994). Glacier physics (particularly thermal regime, structure and hydrology) exerts a fundamental control on many aspects of subglacial eruptions. It is used below to construct three simple models for subglacial eruptions. Comprehensive recent reviews of glacier hydrology and other aspects of glacier physics relevant to this paper have been given by Paterson (1994) and Menzies (1995). The present discussion is not intended to be exhaustive and a fuller treatment will be presented elsewhere. Although empirical, the general principles can be applied to many local situations and they may be able to explain satisfactorily the principal features of most subglacial eruptions.

Glacier thermal regime

Glaciers or parts of glaciers (zones) can be classified according to temperature distribution into two broad categories, temperate and polar. A temperate glacier is at the melting point throughout (although periods of melting alternate with refreezing), whereas temperatures in polar glaciers are well below freezing throughout the year and they are frozen to their bed. This classification is oversimplified and conditions in most glaciers are dynamic and vary from one point to another, both spatially and temporally. Despite this variation, it is useful to conduct the ensuing discussion as if only two categories of glacier thermal regime existed. The discussion is further limited to temperate glaciers (or cold glaciers in which melting point is reached at the bed) because water can migrate under these conditions, a condition that is most informative for subglacial eruptions.

Glacier hydrology

Large quantities of meltwater are created during subglacial eruptions, and glacier hydrology exerts a dominant influence on the sequence of events in such eruptions and on the distribution and types of lithofacies formed. In turn, glacier hydrology is controlled to a large extent by glacier structure, itself mainly determined by snow densification (see next section). Glacier hydraulics describes the motion of water within a glacier system. The most important hydraulic effect of snow densification is on glacier permeability. In simple terms, snow and firn are aquifers and ice is an aquiclude (illustrated graphically by Gore, 1992, fig. 2). Crevasses are also aquifers. Provided the ice is temperate, englacial seepage may also occur in unfractured ice along crystal boundaries or via a network of connected small tubes and veins but, because of the impermeability of ice, hydraulic flow by such seepage is almost negligible compared with flow along the underlying bedrock surface. Meltwater can also pass through an underlying permeable sediment layer (Darcian flow) but, in most instances, meltwater will flow mainly along the glacier bed.

Any discussion of glacier hydrology must focus on the hydraulic gradient or gradient of the water pressure potential of a system, which determines the flow direction of any meltwater. The hydraulic gradient within ice, at the glacier bed and in any underlying permeable sediment layer is controlled predominantly by ice surface slope (Björnsson, 1988; Syverson *et al.*, 1994). Consequently, for many glaciers, water flows in the general direction of the surface slope. During eruptions, the ice surface subsides as ice is melted over the eruption site. Initially, water may drain subglacially but distortion of the local hydraulic gradient by the depressed ice surface rapidly constrains meltwater to flow towards the eruption site. Thus the site becomes effectively sealed by an encircling ice barrier, and a water-filled vault forms above the volcano, melting to the surface and forming an englacial lake (Björnsson, 1988). The lake will drain subglacially when the water becomes deep enough to float the ice barrier, resulting in catastrophic floods (known as jökulhlaups in Iceland).

Glacier structure

After precipitation, snow is transformed into ice by a process of densification that involves the collapse of the snow particle lattice framework and the progressive compression and elimination of trapped air. The transformation and its rate depend on temperature. It happens much more rapidly in glaciers in temperate regions compared with polar regions. Densification results in glaciers divided into different zones or layers, corresponding to snow, firn and ice, which differ from each other in their temperature and other physical characteristics. 'Snow' refers to particulate ice crystals that are essentially unchanged since they were precipitated; 'ice' comprises ice crystals and grains in which the interconnecting air passages have been sealed off; and 'firn' refers to the intermediate stages of transformation of snow into ice. Fundamental changes in physical properties occur during densification. Typical values for density are in the range of 50–400 kg m^{-3} for snow, 400–800 kg m^{-3} for firn, and 800–917 kg m^{-3} for glacier ice (Paterson, 1994).

Fractures (crevasses) form in areas where glacier ice is undergoing extension (e.g. overlying subglacial topographic highs), where tensile stresses exceed the tensile strength of ice. However, crevasses can remain open down to depths of only ≈ 20–30 m as internal deformation pressure acts to seal them at greater depths. This statement ignores the influence of meltwater in crevasses, which can cause fractures to remain open to greater depths. Thus, a layer of crevassed ice may also be present above unfractured ice in some glaciers. For simplicity, the discussion that follows will consider a layered glacier structure, comprising an upper 'firn' layer (including both snow and firn), an intermediate layer of crevassed ice and a lower layer of massive ice. The broad use of the term 'firn' follows the convention of Paterson (1994). It has no practical effect on the conclusions reached below. For purposes of discussion, the presence and hydraulic effects of dirt bands and fold structures caused by glacier flow and deformation over uneven bedrock are also ignored in favour of a simple layered sheet of glacier ice overlying an even horizontal or gently sloping bedrock (Fig. 4).

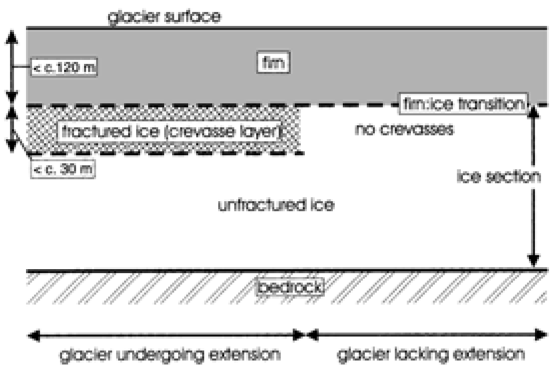

Fig. 4. Section through an idealized glacier showing the empirical layered structure used in this paper to model the hydrodynamic effects of subglacial eruptions. Although only temperate glaciers are considered in this paper, polar glaciers will show a similar structure. Not to scale.

Effects of thermal regime, glacier structure and hydrology on subglacial eruptions

During eruptions beneath glaciers, distinctive sequences of lithofacies are formed whose lithofacies types and architecture are controlled by the thickness and structure of the overlying glacier (Smellie *et al.*, 1993a; Smellie & Skilling, 1994). Thus far, two broad types of eruptive sequence have been documented, corresponding to eruptions beneath 'thick' and 'thin' glaciers. Beneath 'thick' glaciers (i.e. greater than ≈ 200 m thick), a basal section of pillow lavas and/or subaqueous volcaniclastic lithofacies (mainly syneruptive redeposited tephra; McPhie *et al.*, 1993) is formed in a water-filled vault or lake and is overlain by subaerial lavas and cogenetic breccias of hyaloclastite delta(s) (Jones, 1969, 1970; Skilling, 1994; Smellie & Skilling, 1994; Smellie & Hole, 1997). This sequence type and its lithofacies are characteristic of subglacially erupted table-mountain (tuya) volcanoes in Iceland and elsewhere, and they can be difficult to distinguish from lacustrine or marine volcanoes (see Staudigel & Schmincke, 1984; Werner *et al.*, 1996; White, 1996). However, the common draping of subaqueous sections by subaerial lithofacies in the products of a single eruption is one important way in which englacial volcanic successions can be distinguished from other subaqueous volcanic successions (Wörner & Viereck, 1987; Smellie *et al.*, 1993a; Skilling, 1994; Smellie & Skilling, 1994; Smellie & Hole, 1997). Conversely, a vault or lake cannot form beneath 'thin' glaciers (i.e. composed of firn and/or crevassed ice; typically < 100 m thick). Although in theory water may be directed initially into the eruption site owing to distortion of the local hydraulic gradient, a sealed vault cannot be sustained. Thermal erosion of the firn layer by viscous heat dissipation from the meltwater will result in the meltwater rapidly becoming connected hydraulically with the rest of the glacier and flushing away. Open fractures in glacier ice will also allow water to escape. In both circumstances, any meltwater will be removed continuously at the glacier bed during the course of an eruption, resulting in a distinctive association of volcanic and volcaniclastic lithofacies that are characterized by evidence for flowing rather than ponded water, i.e. of 'subglacial sheet-flow type' (model 1, illustrated in Fig. 5a; Smellie *et al.*, 1993a). Measured depths to the firn–ice transition range between 38 and 115 m, although 60–70 m is typical (Paterson, 1994, table 2.2). With a crevasse layer present (≤ 30 m thick), this gives an approximate maximum limit of 150 to perhaps 200 m of firn and crevasses for eruptions yielding such volcanic sequences.

The common occurrence of subglacially erupted table-mountain (tuya) volcanoes has been explained by a hydraulic theory in which the volcanoes are constructed within englacial water-filled vaults or lakes (Björnsson, 1988). Eruptions are initially subaqueous, becoming subaerial when the vents rise above the lake surface. The presence of laterally extensive hyaloclastite deltas implies a period of relative stability of the encircling glacial lake, with which subaerial lavas interacted and were shattered into hyaloclastite delta foreset beds along a passage zone. Meltwater is envisaged to accumulate until the hydrostatic pressure at the base of the vault (P_v) exactly equals that beneath the surrounding impermeable ice barrier (P_i). The barrier is then floated and the meltwater drains catastrophically, resulting in volcanic floods (jökulhlaups).

However, a geological problem exists in applying the conventional hydraulic theory. There is ample magmatic heat in the system to melt large volumes of ice (Allen, 1980; Gudmundsson & Björnsson, 1991). However, in a closed system, at least 10 units of ice need to be melted for every unit of magma erupted in order to create space for the magma, a consequence of the different relative densities of ice and water. Although, theoretically, additional space for meltwater could be created by doming of the ice early in the eruption, there are no observations of historical eruptions to confirm whether doming is a common effect. Moreover, melting is typically so rapid (e.g. several hundred metres of ice were penetrated by melting in only a few hours during the 1996 subglacial eruption in Iceland; Gudmundsson *et al.*, 1997) that any domed ice carapace will rapidly disintegrate over the vent site. Thus, the volume of meltwater must increase at a much faster rate than magma is emplaced. Because of the disproportionate volumes involved, P_v will equal P_i long before subaerial emergence of the volcano can take place. Important geological consequences of this observation for the construction of englacial volcanoes are that englacial vaults will very rapidly become lakes and the eruptive vents must be submerged for much of their history, becoming subaerial only when specific hydraulic conditions are met. For example, when $P_v = P_i$, the barrier will float and the vault will drain subglacially, thus exposing the vent (model 2, illustrated in Fig. 5b). If the eruption continues, the subaqueous lithofacies will be juxtaposed with subaerial lithofacies emplaced in the empty vault. However, in draining, P_v becomes much smaller than P_i, thus allowing the ice barrier to reinstate. The vault may then refill and the process is set to repeat. The final volcano may have an extremely complex lithofacies architecture (implicit in the fledgling models of Smellie *et al.*, 1993a, fig. 14). It is hard to see how a period of stable water level and hyaloclastite delta progradation could occur under conditions where the ice barrier can be lifted and subglacial drainage occur.

By contrast, most natural glaciers (> 100 m thick) have a layered structure, comprising (from base up) massive ice, fractured ice and/or firn (Fig. 4). This feature was not considered by previous workers, and glaciers were depicted as homogeneous ice. In model 2, the overlying permeable layer(s) was too thin relative to the underlying impermeable ice and floating of the ice barrier was an inevitable consequence of $P_v = P_i$. However, flotation cannot occur if the meltwater surface in the vault intersects an aquifer while $P_v < P_i$. As in eruptions beneath thin glaciers, meltwater will escape through the aquifer and the vault will drain by overflowing (model 3, illustrated in Fig. 5c). In this situation, the surface of the englacial lake can be relatively stable during the eruption, with the possibility of forming laterally extensive hyaloclastite deltas during subaerial effusion. In practice, however, water levels are likely to diminish slowly owing to thermal erosion at the outflow point(s), and this may lead to the coeval subaqueous lithofacies being draped by subaerial lithofacies. Overflowing may be the commonest situation responsible for most table-mountain volcanoes with extensive hyaloclastite deltas, and this is another specific hydraulic situation in which the (initially submerged) vents in englacial volcanoes can become subaerial. Calculations suggest that overflowing is likely

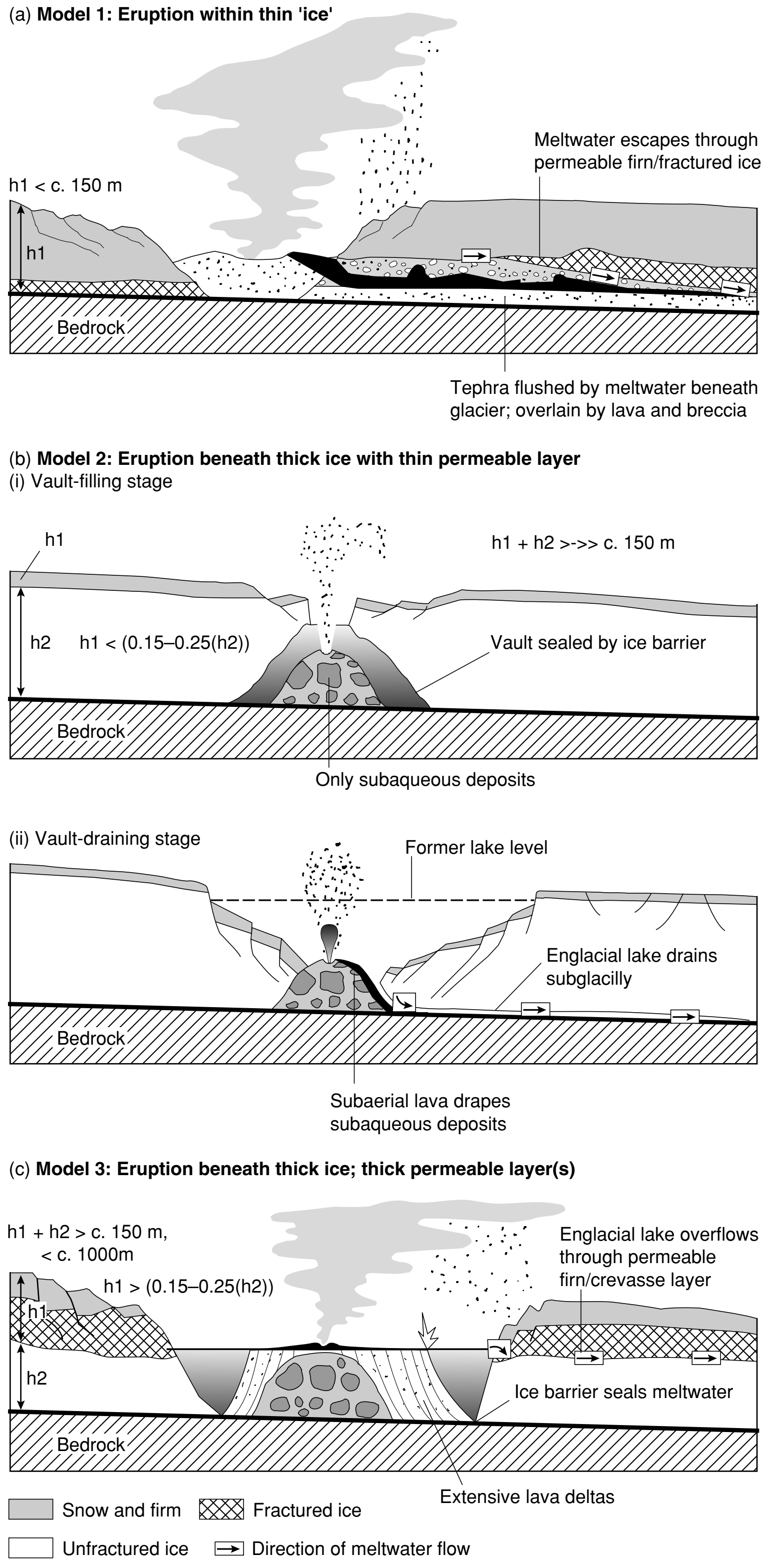

Fig. 5. Three empirical models for subglacial eruptions beneath glaciers of different thickness. The hydrodynamic conditions are different in each case, resulting in significantly different lithofacies architectures. h1, thickness of firn and/or fractured ice (i.e. permeable layer(s)); h2, thickness of unfractured ice. (See text for further explanation.) Not to scale.

to occur only where firn and/or fractured ice layers are greater than ≈ 10–25% of the thickness of unfractured ice. Moreover, influence of firn and fractured ice is negligible in glaciers > 1000 m thick and meltwater generated in these situations is likely to drain only subglacially (i.e. model 2). Overflowing was observed for the first time above a subglacial eruptive fissure during the October 1996 eruption in Vatnajokull, Iceland (Gudmundsson *et al.*, 1997), although concurrent subglacial drainage (through subglacial conduits maintained by thermal erosion) and the short duration of the eruption prevented a lava delta forming.

LITHOFACIES

Icefall Nunatak consists of subaqueously emplaced lavas and volcaniclastic lithofacies dominated by sideromelane lapilli and ash (Table 1). The dominant glassy clasts in most lithofacies are angular and have cuspate to blocky shapes unmodified by sedimentary processes. These pyroclasts have shapes and variable vesicularity consistent with explosive hydrovolcanic disruption of a vesiculating magma (Houghton & Wilson, 1989). Moreover, non-juvenile clasts are very uncommon, suggesting that the fragmentation was driven by surface water rather than ground water, as otherwise clasts derived from the aquifer would be more common (White, 1996). The presence of intact and fragmented basalt pillows in the volcaniclastic sequence, together with pillow lava, turbidites, hyaloclastite breccia and water-chilled sheet lavas (see below and Table 1) also indicates a general subaqueous environment of deposition.

The lithofacies have been named using the non-genetic descriptive terminology of McPhie *et al.* (1993). For brevity, the use of 'volcanic' is omitted from the lithofacies names. They were probably formed mainly by syneruptive redeposition of pyroclastic and autoclastic particles. Initial division in the clastic lithofacies is by dominant grain size. They are mainly types of breccia (prefix 'B', Table 1), including stratified and massive varieties. Further subdivision is based on dominant clast type (vesicular and non-vesicular sideromelane (subscripts 'vs', 'nvs'), tachylite ('sc' after dominant scoria clasts), crystalline ('lith' indicating lithic; note that these are almost invariably juvenile clasts)). There are fewer lithofacies of sand grade (prefix 'S') but sandy sediments are volumetrically abundant in parts of the sequence. Muddy lithofacies are minor and always coupled genetically to sandstones (lithofacies SM). Lavas are divided into blocky-jointed and columnar types (Lj and Lcol, respectively) and pillow lava (Lp). Some basalt units are intrusions and others are transitional to some of the clastic lithofacies as a result of *in situ* brecciation and/or redeposition (e.g. Bnvs, Blith). Disturbed and contorted strata are also distinguished as a separate lithofacies (Z).

VOLCANIC EVOLUTION OF ICEFALL NUNATAK

Icefall Nunatak is a polygenetic edifice constructed from the products of at least five vents (Fig. 1). Three of the vents were located within the outcrop area. Of these, two are occupied by basalt intrusions, whereas the approximate position of the third is identified by vent-proximal lithofacies (welded Strombolian scoria in a cinder cone remnant). The remaining two vents erupted basalt magma with different mineralogical characteristics (see below), but their location is only poorly constrained, using homoclinal to crudely radial bedding orientations and lithofacies characteristics. However, by comparison with other subglacial volcanoes (e.g. Jones, 1969), the two poorly located vents are likely to have been situated < 1 km away from the preserved outcrop.

The construction of the Icefall Nunatak volcano can be described in three major stages of development (I–III), each separated by a prominent unconformity, although stage II was responsible for most of the exposed deposits (Fig. 3; Table 2). Other internal erosive boundaries provide further scope for subdivision within each stage (Ia–c; IIa–h; IIIa–b). Stage I is represented by basalt lavas and breccias, which were emplaced entirely subaqueously, probably mainly in an englacial vault. These are overlain by stage II basalt lavas, breccias and gravelly sandstones, which evolved from a subaqueous (englacial vault, then lake) sequence of vertically aggraded mainly syneruptive redeposited tephra into a lacustrine lava delta having a subaerial lava topset. Stage III is the thinnest sequence present and is largely subaerial. It comprises a thick water-cooled basalt lava and breccia overlain by a small cinder cone relic. The magma erupted throughout the volcano's history was alkali basalt in composition (Table 3). Mineralogical and textural differences are evident between basalt lava erupted in each stage but are consistent within stages. Stage I basalts are essentially aphyric, whereas phenocrysts (mainly olivine, clinopyroxene and plagioclase) are conspicuous in the other two stages. However, phenocrysts are

Table 1. Summary and interpretation of the principal lithofacies observed at Icefall Nunatak

Lithofacies	Code	Description	Interpretation	Unit*
Stratified vesicular sideromelane breccia	Bvs	Sandy gravelly coarse breccia; poorly sorted; planar stratification; often very thick beds (10 cm–3 m); mainly massive or with reverse-graded bases; often normal graded in top 10–20 cm, passing up through 10 cm of faintly planar laminated fine sandstone–siltstone into massive mudstone; 5% conspicuous large clasts (to 80 cm), mainly in massive centres to beds; erosive bases and amalgamation common; large clasts include gravelly sandstone and mudstone up to 50 cm in diameter	As SGp but coarser grained; sedimentation from successive high- and low-density stages in turbidity currents under declining flow conditions; sediment input probably represents freshly erupted resedimented tephra linked directly to eruption columns or jets and slumping; mainly R2R3–TbTe turbidites of Lowe (1982)	Ic; IIc
Stratified non-vesicular sideromelane breccia	Bnvs	Gravelly breccia, minor coarse sandstone; fines poor; indistinct planar stratification dipping at up to 30°; numerous irregular pillows, often flattened, up to 80 cm across, with wrinkled and chilled glassy surfaces; stratification may be lens-like down-dip on a decametre scale	Sideromelane clasts generated mainly by spalling at the topset–foreset 'brinkpoint' in a hyaloclastite (lava) delta; emplaced by either a continuous 'rain' of particulate debris avalanching down the steep delta foreset slope or by redeposition from density-modified grain flows; essentially hyaloclastite breccia	IIg, IIh
Massive sideromelane breccia	Bm	Grey fines-poor or fines-free gravelly breccia; mainly massive; very rare indistinct and local planar stratification and one dune bedform seen; possible rare segregation pipes; mixed vesicular and (mainly) poor to non-vesicular sideromelane; locally rich in fine-grained crystalline clasts; dispersed basalt pillows with glassy rinds; drapes and intrudes fractures in units IIc and IId	Uncertain sedimentological interpretation but field relations suggest deposit followed major slope failure event; probably slump deposit emplaced as large cohesionless debris flow	IIe
Massive to poorly stratified lithic breccia	Blith	Matrix-poor or matrix-free gravelly breccia; dominated by fine-grained crystalline clasts; blocky shapes, non-vesicular; minor sandy matrix of sideromelane clasts variably crowded with crystallites; minor tachylite; minor complete and fragmented basalt pillows; grades laterally into blocky-jointed lava (Lj); mainly massive; rare faint stratification and planar beds; local jigsaw breccia	Clasts generated by mechanical rupture during subaqueous lava emplacement; many clasts formed *in situ*, others underwent minor redeposition probably mainly by local avalanching in density modified grain flows	Ia, Ib, Ic; IIf, IIh; IIIa?
Tachylite scoria breccia	Bsc	Gravelly matrix-poor breccia bed up to 6 m thick; grey, maroon and khaki; formed of moderate to highly vesicular, rarely incipiently vesicular lapilli, many with fluidal surfaces; blocky shapes and large ragged bombs also common; variably tachylite- or sideromelane-rich; local superficial palagonite alteration and pervasive zeolite (unit Ib); As above, plus poor discontinuous planar stratification; dense ovoid bombs common, some cored; locally welded; few thin (10–40 cm) sand-grade beds, reverse-graded to massive, traceable over metres because of yellow coloration, contain abundant sideromelane droplets (achneliths) (unit IIIb)	Proximal pyroclastic deposits (lapillistones) of Strombolian eruptions; subaerial cinder cone remnant (unit IIIb) and subaqueous stratum (Ib); palagonite and zeolite alteration in IIIb as a result of structural position of cone, on top of water-saturated volcanic edifice associated with dyke intrusion; pyroclasts in Ib possibly formed within steam cupola surrounding subaqueous eruption column, thus reducing contact with water until after deposition	Ib; IIIb
Planar-stratified gravelly sandstone	SGp	Gravelly coarse sandstone to sandy fine breccia; clasts typically 1 mm to 2 cm, mainly sideromelane, variably vesicular; planar continuous stratification; beds 5 cm to a few metres thick; few	Subaqueously and subaerially erupted hydroclastic tephra redeposited by cohesionless debris flows or high-density turbidity currents [(S1,S2)S3 turbidites of Lowe, 1982]	Ic; IIb, IIIc, IId

		large lithic clasts (to 40 cm, typically < 25 cm); poorly to well sorted; mostly massive or normal graded; rare reverse-graded bases; rare planar lamination; erosive bases and amalgamation common; few load structures, some containing ballistic blocks; very common lithofacies		
Cross-bedded gravelly sandstone	SGt	Fine sideromelane breccia to medium sandstone; dominated by traction current bedforms, mainly planar and trough cross-stratification; beds lens-like, none traceable > 5–10 m; bed thicknesses 10–40 cm	Syneruptive reworking and clast-by-clast deposition from traction currents	IIa
Graded–laminated fine sandstone–mudstone	SM	Thin planar beds (1–5 cm thick); form discrete sequences a few dm thick (up to 1.8 m); sandstones have sharp bases, normal grading and small-scale load structures; some climbing ripple cross-sets; pass up gradationally into massive dark brown mudstone	Subaqueously and subaerially erupted fine hydroclastic tephra redeposited as Tae, Tbe and Tce turbidites from low-density turbidity currents; some may represent residual flows which were completely detached from, and bypassed, their high-density turbidity current precursors	IIc
Blocky-jointed massive lava	Lj	< 1–20 m thick; very irregular pods, lenses and more continuous basalt lavas characterized by blocky to hackly jointing; joint surfaces commonly rust-stained; may change laterally into columnar or pillow lava, or monomict lithic breccia	Basalt lava extruded into water and intruded into water-saturated sediments; the jointing is a consequence of rapid cooling in contact with water [cf. 'kubbaberg' (box-jointed) lavas in Iceland described by Bergh & Sigvaldason, 1991]	Ia, Ib; IIf, IIg, IIh
Columnar-jointed lava	Lcol	Forms irregular parts of some blocky-jointed lavas (Lj) and a 20-m-thick lava at the crest of the nunatak; the latter contains a thin (2–4 m) basal colonnade overlain by a well-developed entablature; traced laterally, the lava changes into a sharply cross-cutting columnar intrusion	Columnar basalt is common in subaerial sequences but association with sideromelane breccia (Bnvs) and blocky-jointed lava (Lj) indicates subaqueous environment probably corresponding to passage zone in hyaloclastite (lava) delta; suppressed colonnade and expanded entablature in upper 20-m-thick lava indicate rapid cooling of a lava flooded by water; upper lava transforms to a neck at west end of outcrop	(IIf, IIh); IIIa
Pillow lava	Lp	Isolated, complete and fragmented basalt pillows and irregular lens-like masses of pillow lava; locally common; individual pillows are up to 80 cm across, ovoid to flattened, with chilled glassy rinds up to 1 cm thick and rare wrinkled surfaces; no interpillow sediment	Pillow lava and pillow intrusion in wet sediments	Ia, (Ic); (IIe), IIf, (IIg)
Contorted or disturbed strata	Z	Z1: Heterogeneous mixture of sandstone clasts (dm to rarely 5 m diam.) dispersed in massive sandstone matrix; 'swirling' textures common in latter; clasts have diffuse rounded to relatively sharp angular margins; diffuse dewatering structures common; penetrative shear-like joint fabric rarely present; forms steep cross-cutting zones spatially associated with faults and/or upper surface of unit IIc	Zones of focused dewatering possibly induced by syn-sedimentary faulting and/or pressure release associated with sector collapse of tuff cone flanks; essentially formed *in situ*; rare shear-like fabric possibly induced in more consolidated sediments by faulting and/or sliding of large blocks (lithofacies Z3)	IIc
		Z2: Folded and 'slurry-like' beds up to 1.5 m thick; the latter closely resemble Z1 in appearance but always contain dark brown sandy mudstone matrix and are conformable beds	Folding as a result of synsedimentary gravity-induced sliding that may have generated slump sheets; slurry-like beds formed by density inversion (Rayleigh–Taylor instabilities) leading to gravitational collapse and mixing of sandstone and mudstone strata essentially *in situ*	IIc
		Z3: large (up to 20 m diam.) blocks of stratified sediment (mainly SGp) with variable often steeply dipping orientations; corresponds to unit IId	Gravity-induced detachment of blocks of indurated strata and translation down the tuff cone flanks	IId

*See Table 2 and Fig. 3 for description and distribution of units.

Table 2. Summary of evolutionary stages, units and characteristic lithofacies of volcanoes at Icefall Nunatak

Stage	Description of main event(s)	Units*	Principal lithofacies†
III	Lava effusion into thin ice; Strombolian eruptions	a, b	Tachylite scoria breccia (Bsc), columnar-jointed lava (Lcol); possible massive lithic breccia (Blith)
II	Lateral extension by hyaloclastite	f–h	Stratified non-vesicular sideromelane breccia (Bnvs); blocky-jointed lava (Lj); (lava) deltas; lithic breccia (Blith)
	Subaerial emergence, slope failure (faulting, sector collapse)	d, e	Massive sideromelane breccia (Bm); large rotated blocks of SGp, slurry-like and contorted beds (Z)
	Subaqueous tuff cone (rapid planar-stratified gravelly vertical aggradation)	b, c	Stratified vesicular sideromelane breccia (Bvs); sandstone (SGp); graded–laminated sandstone–mudstone (SM)
	Subglacial sediment flushing	a	Cross-bedded gravelly sandstone (SGt)
I	Subaqueous tuff cone (in part)	c	Stratified sideromelane breccia (Bvs); resedimented lithic breccia (Blith)
	Subglacial 'pillow volcano'	a, b	Blocky-jointed lava (Lj), pillow lava (Lp), lithic breccia (Blith)

*See Fig. 3 for distribution of units.
†See Table 1 for lithofacies codes and descriptions.

generally more abundant and larger in stage III basalts, and they also have more coarsely crystalline groundmasses than in preceding stages. By contrast, compositional differences are slight and can be explained by variable accumulation of phenocrysts (particularly ferromagnesian minerals).

The three growth stages can be interpreted as products of a single continuous volcanic growth cycle. Alternatively, the multiple vents involved, and evidence for breaks in the succession (unconformities and petrological differences between stages), raise the possibility that the three major parts of the succession are from unrelated vents active at very different times. However, the lack of significant geochemical differences between stages, and the age dating constraints (indicating a relatively short eruptive period), suggest that any breaks in eruption were probably short (sufficient for only minor fractional crystallization, but sufficient to create differences in the size and relative abundances of the erupted crystals). A genetic connection between the three stages is more likely.

Stage I ('pillow volcano' stage; englacial vault; possible transition to subaqueous tuff cone)

Description

Unit I is restricted to the eastern basal part of the outcrop. It has an exposed maximum thickness of ≈ 60 m. Three subunits are recognized, and are mainly formed of monomict lithic breccia and blocky-jointed lava.

Unit Ia has a sloping upper contact dipping southwest at about 30° (Figs 3 & 6). It comprises 50–60 m of (juvenile) lithic orthobreccia (Blith), which is predominantly massive but locally shows faint crude stratification dipping parallel to the upper contact. The angular basalt clasts range from fine gravel to cobbles 5–10 cm in diameter and there are dispersed basalt pillows. The unit is coarser downward, where it contains irregular lenses and lobes of blocky-jointed and pillowed lava (Lj, Lp) a few metres thick (Fig. 7). Small parts of the deposit are rich in non-vesicular sideromelane clasts and lava pillows. Sideromelane clasts rich in small crystals (commonly 50 vol. %) are ubiquitous but generally minor volumetrically, restricted to the gravelly sandy size fraction, and the majority of clasts are holocrystalline (lithic (juvenile)); zeolite and/or carbonate cement is locally conspicuous. Unit Ib closely resembles Ia, comprising ≈ 12 m of basal blocky-jointed lava intimately associated with irregular patches of rust-stained lithic breccia. It is overlain by 20 m of massive gravelly breccia with lenses of scoriaceous and non-vesicular lava. The shapes of some of the lava units are very irregular, and large and small lava lobes protrude into the enclosing breccia. The lavas and breccias are overlain by two thin (1–6 m) basalt sheet lavas separated by ≈ 4 m of crudely planar bedded, massive gravelly breccias. The two upper lavas have scoriaceous surfaces and parts of the uppermost lava are entirely scoriaceous. It transforms westwards into coarse scoria breccia (Bsc), and is possibly continuous with a similar vesicular

Table 3. Chemical compositions and summary petrography of lavas at Icefall Nunatak

Sample:	MB.48.10	MB.48.5	MB.48.21	MB.48.15
$^{40}Ar/^{39}Ar$ age (Ma):	6.80 ± 0.10		6.47 ± 0.08	
Unit:	Ib	IIf	IIIa	IIIb
Lithology:	alk. basalt lava	alk. basalt lava	alk. basalt lava	alk. basalt bomb
Summary petrography:	< 5% phenocrysts of ol > cpx >> pl	25% phenocrysts of ol = pl >> cpx >> sp	25–40% phenocrysts of ol = pl >> cpx >> sp > ap	30% phenocrysts of ol >> cpx >> pl >> sp
	Crystals to 0.6 mm	Crystals to 3 mm	Crystals to 5 mm	Crystals to 3.5 mm
	Fine groundmass	Fine groundmass	Coarse groundmass	Coarse groundmass
SiO_2	44.46	46.37	45.76	44.77
TiO_2	2.78	2.47	2.69	2.85
Al_2O_3	15.27	15.29	15.36	15.17
$Fe_2O_3^T$	12.90	12.24	13.25	12.96
MnO	0.18	0.24	0.20	0.18
MgO	7.12	5.88	8.94	8.19
CaO	11.45	9.83	9.33	10.28
Na_2O	3.04	3.34	3.28	3.29
K_2O	1.04	1.10	0.84	0.72
P_2O_5	0.64	0.57	0.71	0.69
LOI	1.59	2.83	–0.03	0.79
Total	100.48	100.16	100.31	99.88
Normative Ol	13.92	13.48	20.46	17.50
Normative Ne	5.97	1.81	2.70	4.41
mg-number	0.52	0.49	0.57	0.56
Cr	195	317	225	268
Ni	85	129	197	137
Cu	73	64	60	68
Zn	94	102	89	94
Ga	22	20	19	20
Rb	18	22	12	9
Sr	672	655	752	728
Y	27	27	27	29
Zr	171	186	165	217
Nb	38	40	37	38
Ba	240	324	286	241
La	20	31	31	23
Ce	72	68	59	77
Nd	29	29	28	34
Pb	8	8	9	10
V	248	203	189	206

All samples analysed by X-ray fluorescence at the University of Keele, UK, using standard procedures. $Fe_2O_3^T$, all iron calculated as Fe_2O_3. LOI, loss on ignition. Normative olivine (Ol) and nepheline (Ne) calculated using $Fe_2O_3/FeO = 0.2$. *mg*-number = $MgO/(FeO^T + MgO)$, using molar proportions, where FeO^T is all iron calculated as FeO.

dyke nearby to the west although the exposures are discontinuous.

Unit Ic is > 30 m thick. It is conformable on top of unit Ib and wedges out against the sloping surface of unit Ia. Beds within Ic terminate against Ia without thinning. The unit comprises crudely stratified, blocky and gravelly breccia beds and lenses (Blith) with dispersed basalt pillows up to 80 cm across, alternating with continuous beds of gravelly sandstone (SGp) 30 cm to 4 m thick. The latter have erosive bases, commonly with a thin basal reverse-graded interval and thin normal-graded tops. They have abundant tachylite and sideromelane clasts with a wide range of vesicularities (incipient to high; categories after Houghton & Wilson, 1989) together with highly crystalline non-vesicular grains resembling clasts in the lithic breccias (Blith). By comparison with units Ia and Ib, the sideromelane and tachylite clasts in Ic are crystal poor (0–5%, rising to ≈ 15% in tachylite).

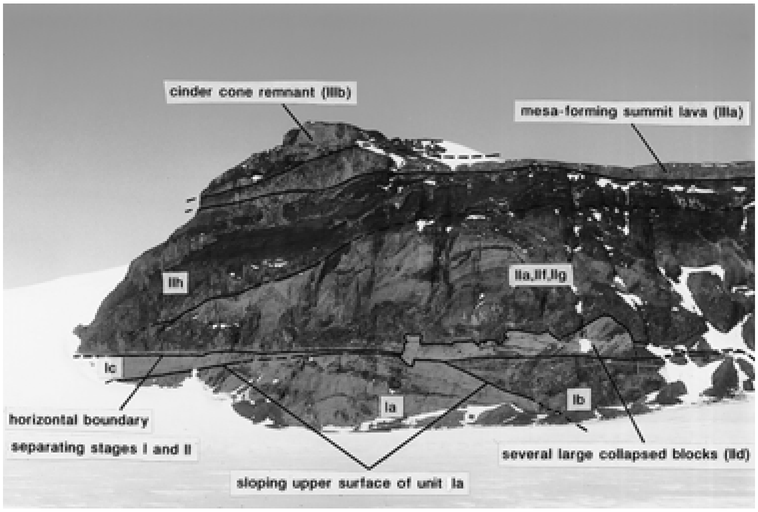

Fig. 6. View of the 200-m-high rock section exposed at the east end of Icefall Nunatak. The cliff sequences were constructed during three principal eruptive stages described in this paper. Particularly prominent features include a subhorizontal surface separating stages I and II, several very large collapsed stage II blocks (lithofacies Z3, unit IId), and homoclinally sloping large-scale breccia beds and interbedded lavas of stage II lava delta(s) (units IIg, IIh). A single thick basalt lava (unit IIIa) forms the cliff edge at upper right side and the highest exposures are formed by a cinder cone relict (unit IIIb). (Note that the viewpoint for this photograph differs slightly from that used to construct Fig. 3, hence the details do not correspond exactly.)

Fig. 7. Blocky-jointed basalt lava (Lj) and cogenetic carbonate-cemented, coarse, fines-free lithic breccia (Blith) in unit Ia. The pencil is about 15 cm long.

Interpretation

In a glacial context, this lithofacies formed within an englacial vault and/or possibly lake (Fig. 8) and corresponds structurally, but not in lithofacies, to the predominantly effusive pillow volcano stage observed in many subaqueous volcanoes (e.g. Staudigel & Schmincke, 1984; Moore, 1985; Skilling, 1994; Smellie & Skilling, 1994; Smellie & Hole, 1997). Stage I is dominated by a large thickness of lithic orthobreccia and lesser blocky-jointed lava. The patchy sideromelane matrix in the breccias (Blith) formed from a basalt lava by thermal shock and spalling in a subaqueous environment, but the abundance of lithic (juvenile) fragments and crystal-rich sideromelane indicates derivation mainly from a well-crystallized lava. Juvenile clasts in younger sedimentary beds (unit Ic) are crystal poor, suggesting that the high crystal content of clasts in the lithic breccias was achieved either by crystallization during effusion of the cogenetic basalt lavas or that the lavas were erupted from a separate batch of highly crystallized magma from a vent different from that responsible for the younger units. The field relations and similar petrographical characteristics of the lavas and breccia clasts indicate a cogenetic relationship. The overwhelming dominance of juvenile lithic clasts indicates that they are not hyaloclastite breccias *sensu stricto* but probably formed as joint-block deposits by mechanical breakage analogous to autoclastic brecciation (Kokelaar, 1986). The disruption was probably aided by hydrofracturing and steam explosions during subaqueous