Pesquisa Antártica Brasileira (2004) 4: 67-76
(Brazilian Antarctic Research)
ISSN 0103-4049
www.cnpq.br/areas/terra_meioambiente/proantar/

Bidimensional numerical simulation of the Lange Glacier, King George Island, Antarctica: preliminary results

HELOISA H.C. BARBOZA[1,2], ÁLVARO L. DE BORTOLI[1], JEFFERSON C. SIMÕES[2], RUDNEI D. DA CUNHA[1] and MATTHIAS BRAUN[3]

[1]Programa de Pós-Graduação em Matemática Aplicada, Instituto de Matemática
Universidade Federal do Rio Grande do Sul, Av. Bento Gonçalves, 9500
91501-970 Porto Alegre, RS, Brasil
[2]Núcleo de Pesquisas Antárticas e Glaciológicas, Departamento de Geografia
Instituto de Geociências, Universidade Federal do Rio Grande do Sul, Av. Bento Gonçalves, 9500
91501-970 Porto Alegre, RS, Brasil
[3]Zentrum für Fernerkundung der Landoberfläche, Universität Bonn
Walter-Flex-Str. 3, D-53113 Bonn, Germany

ABSTRACT

A simplified numeric model simulates the flow of the outlet Lange Glacier (King George Island, Antarctica) to estimate its equilibrium state and morphological responses to accumulation rate changes. The bidimensional model, representative and stable, uses the finite difference method to provide qualitative information on superficial velocity. Fieldwork, carried out in the austral summer of 1999/2000, during the XVIII Brazilian Antarctic Operation, provided glacier ice velocity data for calibration. The model indicates that variations in the net accumulation rate, of less than 20% of the present value, do not significantly change the superficial glacier morphology. Even a reduction of 50% of this rate will lower the surface by only 26 m in 100 years. These results suggest a glacier near the 'steady state'.

Key words: ice masses, glacial flow, numerical simulation.

INTRODUCTION

For the last 45 years, several glacier drainage basins have retreated on King George Island (KGI), Antarctica, located between 57°35'S–59°02'W and 61°54'W–62°16'S (Figure 1). Since 1956, this island has lost about 7% of its original ice cover (Simões et al. 1999). This ice loss was concomitant to an estimated mean atmospheric temperature increase of 1.1°C (Ferron et al. 2004, this volume), although the role of this warming on the ice retreat has not been proved. Furthermore, retreating glaciers have complex topographies, including floating ice fronts, which complicate any

Correspondence to: Heloisa H.C. Barboza
E-mail: hhcb@terra.com.br

environmental interpretation about recent morphological changes. It is necessary, therefore, to proceed with further studies to check the reasons of these modifications. In this paper, we use a simple ice flow model to calculate the internal deformation of Lange Glacier, one of the main outlets of the KGI ice cap, and to obtain information about its state of equilibrium and responses to changes in the net accumulation rate.

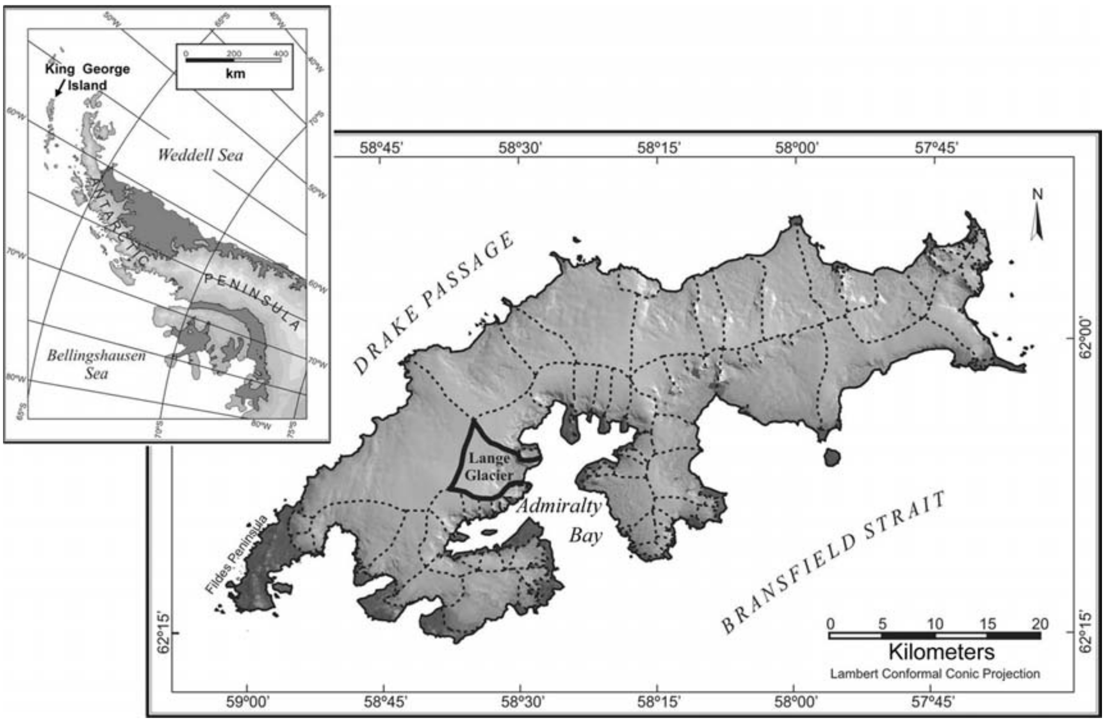

Fig. 1 – Lange Glacier drainage basin location in King George Island (KGI) ice cap. Stippled lines mark drainage basin divides. Inset locates KGI to the west of the Antarctic Peninsula.

THE LANGE GLACIER

Lange Glacier, a main outlet of the KGI ice cap, flows towards Admiralty Bay (Figure 1). Its drainage basin covers 28.3 km^2; it is 6.2 km long across the longitudinal axis, 5.0 km wide in the intermediary portion and its front has a width of about 2 km. The drainage basin was determined using a SPOT satellite image (Simões et al. 1999), radio echo-sounding points (Macheret et al. 1998, Macheret and Moskalevsky 1999), and ice velocity measurements. Field measurements (Simões et al. 2004, this volume) and the presence of sediment plumes in front of its terminus (Pichlmaier et al. 2004, this volume) point to a warm glacier profile. Lange Glacier has had the greatest ice front retreat on KGI since 1956 (Simões et al. 1999), a 1.4 km withdrawal, losing a total of 2.0 km^2 (Arigony-Neto 2001, unpublished).

DATA SOURCES

During the XVIII Brazilian Antarctic Expedition (austral summer of 1999/2000), 21 stakes were fixed on the Lange Glacier surface, in order to measure the ice velocity on the glacier's surface. Two DGPS (Differential Global Position System) with centimetric precision (MAGELLAN PRO MARK 100) were used (Braun et al. 2001), one was kept in a fixed position as a reference, and the other was moved among the stakes to measure their displacement. Depending on measuring time and satellite configuration, measurement errors were in the 5 cm range. Additionally, errors due to melting/compactness of the superficial snow cover or stake relocation, due to melting, were estimated to be less than 10 cm.

The position of these stakes and their velocity vectors are shown in Figure 2. Results confirmed the theoretical expectations. For example, stakes next to the drainage basin limits, determined from the satellite image, move slower than those at the glacier's center. Braun et al. (2001) provides further details on the methodology of these measurements.

The superficial topography, along the 6.2 km profile A-B (Figure 2), from the ice divide to about 400 m above sea level (a.s.l.), was obtained from DGPS surveys carried out in the summers of 1997/98 and 1999/2000 (Braun et al. 2001). Below 400 m of altitude, the topography was derived from the Admiralty Bay map (1: 50,000), published by the Polish Academy of Sciences (Battke 1990). The subglacial profile was obtained using data collected from a radio-echo sounding survey, carried out by Norbert Blindow of the *Universität Münster*, in the summer of 1997/98, and complemented with data from Macheret and Moskalevsky (1999). Figure 3a shows the superficial and subglacial profiles used in the model. Simões et al. (2004, this volume) provided information on the temperature distribution near the ice divide; for each 10 m there was a temperature decrease of 0.3°C.

THE NUMERICAL METHOD

Common methods employed for the solution of fluid/solid deformations are those of finite differences and finite elements. The finite differences method, used in this work, is a simple and efficient method (Anderson et al. 1984).

In order to simulate the ice flow, the Navier equations, with special stress-strain relations, can be employed. However, experimental data show that the velocities of Lange Glacier are sufficiently small, less than 1 m day^{-1}, therefore allowing for some simplifications. In this way, an energy equation is employed to obtain temperature variations. Mass conservation along the domain is based on the cross-area variation.

The average ice velocity through a vertical column, at a point on the ice sheet or an ice cap, is defined by:

$$V = V_i + V_b \tag{1}$$

where V_b is the basal sliding velocity and V_i is the average internal velocity of deformation. In this study, unrealistic as it may appear, we do not calculate V_b, even though the ice-rock interface is

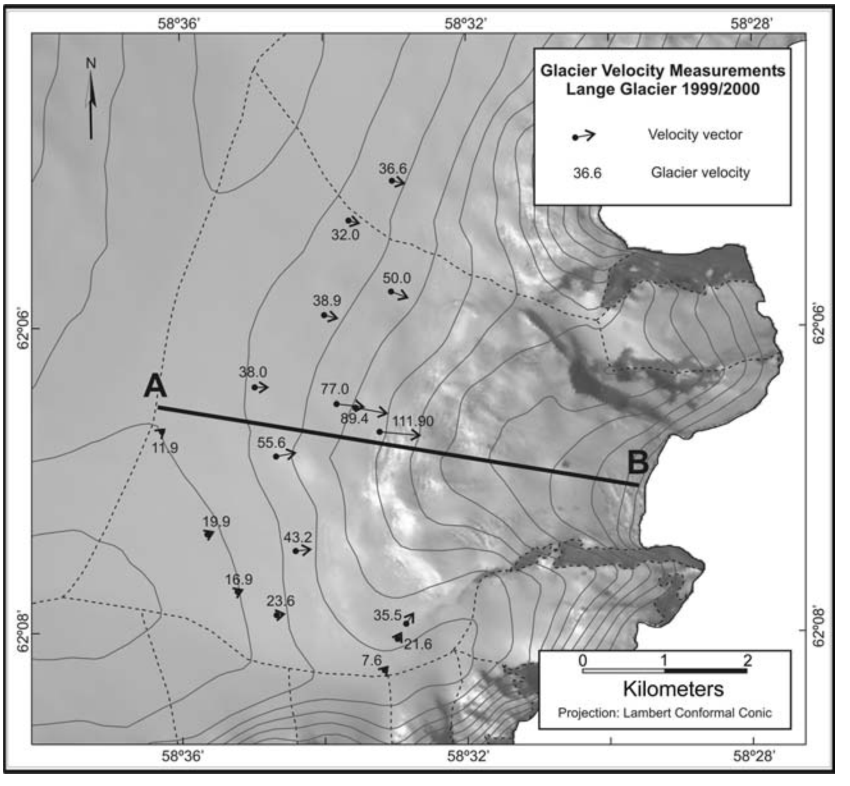

Fig. 2 – SPOT subscene displaying Lange Glacier and its superficial movement pattern with velocity vectors. Line A-B marks the profile modelled in this study.

probably at the melting pressure point. V_i is obtained using an empirical relation (Paterson 1994):

$$V_i = A_1 \tau_b^n h \tag{2}$$

with $n = 3$; A_1 is an empirical parameter, which depends on the ice temperature, crystal size and orientation. We adopted the value $A_1 = 9.512910^{-25}$ Pa, h is the ice thickness; tension τ_b is calculated using equation (3).

$$\tau_b = \rho g \ \alpha \ h \tag{3}$$

where ρ is the average ice density, g is the gravitational acceleration, and α the large-scale surface slope. The convected energy equation is parabolic, enabling the use of central differences approximations for convective as well as diffusive terms (De Bortoli 2000). Therefore, a successive under-relaxation (SUR) scheme is employed with a relaxation coefficient of order 0.8, which is common for incompressible flow situations.

The energy equation (1) for bidimensional flow can be written, in generalized coordinates, in the following way: S contains the source terms, T is the temperature, u the horizontal velocity, α

the thermal diffusivity, t the time and x and y the coordinate directions.

$$\frac{\partial T}{\partial t} + u\left(\left(\frac{\partial T}{\partial_\zeta}\right)\zeta_x + \left(\frac{\partial T}{\partial_\eta}\right)\eta_x\right) = \alpha\left(\left(\frac{\partial^2 T}{\partial_\zeta^2}\right)\zeta_{x^2} + 2\left(\frac{\partial^2 T}{\partial_\zeta \partial_\eta}\right)\zeta_x\eta_x + \left(\frac{\partial^2 T}{\partial_\eta^2}\right)\eta_{x^2} + \right.$$
$$\left. + \left(\frac{\partial^2 T}{\partial_\zeta^2}\right)\zeta_{y^2} + 2\left(\frac{\partial^2 T}{\partial_\zeta \partial_\eta}\right)\zeta_y\eta_y + \left(\frac{\partial^2 T}{\partial_\eta^2} - \right)\eta_{y^2}\right) + S \tag{4}$$

Following the finite differences method, each derivative must be approximated. Forward approximations are used for the time term, while central differences are employed for convective and diffusive terms.

The metrics and Jacobian of the coordinates transformation relate the area and length relations between the physical and computational domains. They are written as follows:

$$\begin{aligned}
\zeta_x &= \tau Y_\eta \\
\zeta_y &= -\tau Y_\zeta \\
\eta_x &= -\tau X_\eta \\
\eta_y &= \tau X_\zeta \\
\tau &= l/(X_\zeta Y_\eta - Y_\zeta X_\eta)
\end{aligned}$$

In order to obtain a numerical solution, boundary conditions for all equations must be established. The appropriate boundary conditions would allow consistent results, taking up a computational of just a few minutes. On the open surface, a Dirichlet condition can be employed, because some experimental data are available. On the other hand, at the bottom surface and at the left and right domain extremities, the boundaries are not completely defined by experiments. Table I lists boundary conditions used in the model.

TABLE I

Boundary conditions used in the numeric model.

	Temperature	Altitude (m a.s.l.)	Ice flux
Surface	−0.3°C	from 84 m to 577 m	U = 1.007 × 10^{-5}
Ice-rock interface	−1.0°C	from 78 m to 256 m	U = ub(i) = u(i,nj)-ui(1)
Left boundary	−1.0°C to −0.3°C	from base to top surface	extrapolation
Right boundary	−1.0°C	extrapolation	extrapolation

Through the continuity equation the change in ice thickness can be expressed (Colbeck 1980). For a two-dimensional model, it results in the following equation:

$$\frac{\partial h}{\partial t} = b(x,t) - \frac{\partial q_x}{\partial x} \tag{5}$$

where b is the accumulation balance and q_x is the ice flux as a function, of its position and time. It is calculated using Equation (6):

$$q_x(x,t) = -2A(n+2)^{-1}(\rho g)^n \alpha^{n-1}\left(\frac{\partial h_s}{\partial x}\right)h^{n+2} + h\, u_b(x,t) \tag{6}$$

where $\alpha = \dfrac{\partial h_s}{\partial x}$ (h_s is the altitude) and $A = 6.8.10^{-15}s^{-1}kPa^{+3}$, using the finite differences method that results in the following equation:

$$q_x(x,t) = -2A(n+2)^{-1}(\rho g)^n \left(\frac{h_s(i+1,j) - h_s(i-1,j)}{2\partial x}\right)^n h^{n+2} + hu_b(x,t) \tag{7}$$

Other important processes, which are known to occur at KGI's glaciers, such as water interacting with the lower boundary (Hambrey 1994) and ice melting, which could make this model very complex, will be investigated in future work.

NUMERICAL RESULTS

The computational mesh for the Lange Glacier contains 155 × 6 equally spaced points. Vertical and horizontal points were established according to available experimental data. Figure 3b shows the computational grid for this flow situation. Figure 4 displays vector fields for small regions around points A and B as indicated in Figure 3a. Velocity tends to increase from bedrock to ice surface and from A to B, as expected.

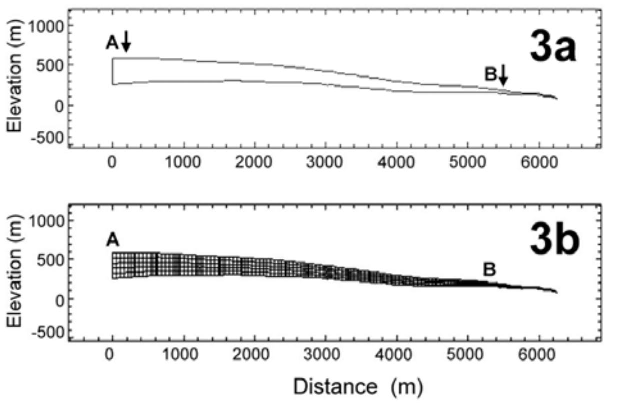

Fig. 3 – Ice surface and bedrock elevation along line A-B (3a), and computational grid for Lange Glacier, 155 × 6 points (3b). Arrows identify areas detailed in Figure 4.

Using the equation of continuity, we observed the response of the Lange Glacier ice when changing the present accumulation rate, approximately 0.6 m y^{-1} water equivalent, see Simões et al. (2004, this volume). Running the model for 100 years, with a net accumulation rate 20% greater than the present one, does not significantly modify the glacier thickness. A net accumulation rate

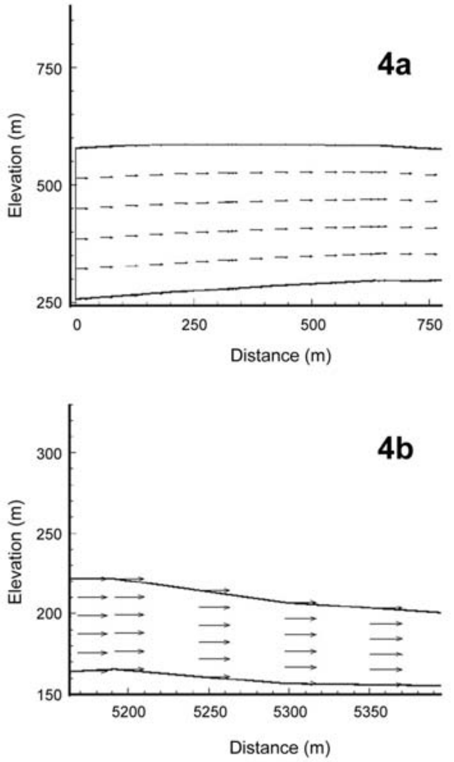

Fig. 4 – Vector fields for regions A (4a) and B (4b), as identified by arrows in Figure 3a.

increase of 50% would result in an increase of about 26 m in the elevation profile (Figure 5). Conversely, a 50% reduction in the net accumulation rate would lower the surface by 26 m, after 100 years (Figure 6).

CONCLUSION

This stable model is able to provide qualitative information on the velocity of the ice column. Small variations in the net accumulation rate, less than 20% of the present value, do not significantly change the surface morphology. The model, when run for 100 years with a reduction of 50% in the net accumulation rate and at a constant temperature, produces a 26 m lowering of the surface before, reaching a new equilibrium state. The Lange Glacier internal deformation is not very sensitive to variations in the net accumulation rate, at least in the temporal scale of this study, less than 100 years. This model does not consider the basal sliding component, therefore, it is unrealistic because the ice that makes up Lange Glacier is at or near the melting pressure point. Further developments must de added to this component and to the impact of surface melting, due to regional atmospheric warming.

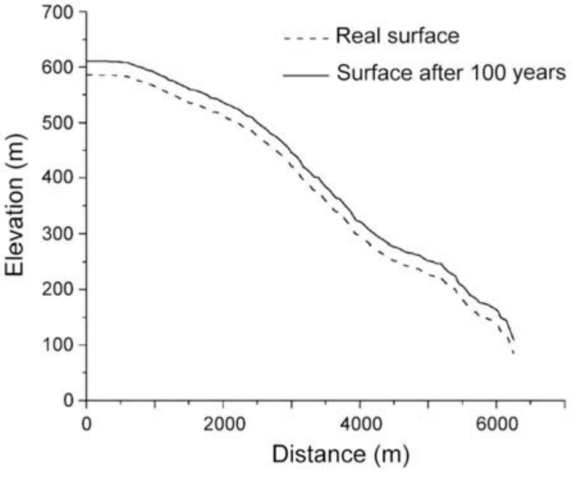

Fig. 5 – Lange glacier surface variation after running the model for 100 years, with a net accumulation rate 50% greater than the present one.

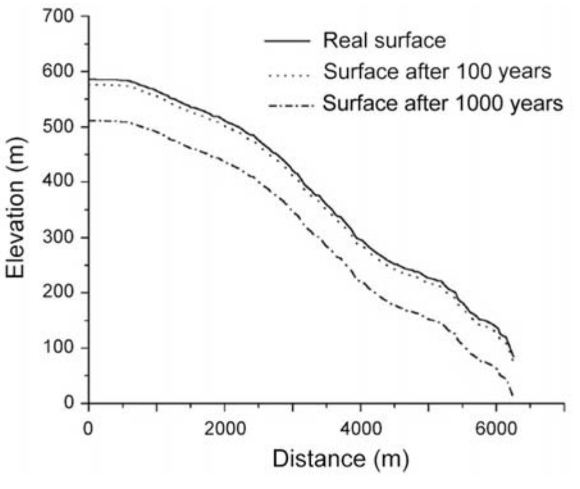

Fig. 6 – Lange glacier surface variation after running the model for 100 and 1000 years with a net accumulation rate 20% smaller than the present one.

ACKNOWLEDGMENTS

The Brazilian Antarctic Programme (PROANTAR), through a grant to JC Simões, Brazilian National Council for Scientific and Technological Development – CNPq project (48.024.3/00-0), supported this work. The first author thanks CNPq for the studentship 190161/99-0. We also thank Norbert Blindow, *Universität Münster*, Germany, for providing the unpublished RES (radio echo-sounding) data.

RESUMO

Neste trabalho, modela-se simplificadamente o fluxo da geleira de descarga Lange (ilha Rei George, Antártica) para estimar seu estado de equilíbrio e as respostas morfológicas às mudanças na taxa de acumulação. O modelo bidimensional, representativo e estável, fornece resultados qualitativos para a velocidade no domínio, baseado no método das diferenças finitas. Para calibrar o modelo, medidas da velocidade da geleira foram obtidas durante a XVIII Operação Antártica Brasileira no verão austral de 1999/2000. O modelo indica que variações na taxa de acumulação líquida, menores que 20% do valor atual, não resultariam em mudanças significativas na morfologia superficial da geleira. Mesmo a redução em 50% desta taxa resultaria somente no rebaixamento da superfície em 26 m, em 100 anos. Estes resultados reforçam a idéia de uma geleira perto do estado de equilíbrio (*steady-state*).

Palavras-chave: massas de gelo, fluxo glacial, simulação numérica.

REFERENCES

ANDERSON DA, TANNEHILL JC AND PLETCHER RH. 1984. Computational Fluid Mechanics and Heat Transfer. New York: McGraw-Hill, 598 p.

ARIGONY-NETO J. 2001. Determinação e interpretação de características glaciológicas e geográficas com sistema de informações geográficas na Área Antártica Especialmente Gerenciada Baía do Almirantado, Ilha Rei George, Antártica. Porto Alegre: Universidade Federal do Rio Grande Sul, 84 p. Unpublished M.Sc. dissertation.

BATTKE Z. 1990. Admiralty Bay, King George Island. Map, 1: 50.000. Warsaw: Nakladem Instytutu Ekologii, Polish Academy of Science.

BRAUN M, SIMÕES JC, VOGT S, BREMER UF, BLINDOW N, PFENDER M, SAURER H, AQUINO FE AND FERRON FA. 2001. An improved topographic database for King George Island – compilation, application and outlook. Antarct Sci 13: 41–52.

DE-BORTOLI AL. 2000. Introdução à Dinâmica de Fluidos Computacional. Porto Alegre: UFRGS, 134 p.

FERRON FA, SIMÕES JC, AQUINO FE AND SETZER AW. 2004. Air temperature time series for King George Island, Antarctica. Pesq Antart Bras 4: 155–169.

HAMBREY MJ. 1994. Glacial Environments. London: UCL Press. 295 p.

MACHERET YUYA AND MOSKALEVSKY MYU. 1999. Study of Lange Glacier on King George Island, Antarctica. Ann Glaciol 29: 202–206.

MACHERET YUYA, MOSKALEVSKY MYU, SIMÕES JC AND LADOUCH L. 1998. Structure and regime of King George Island ice sheet, South Shetlands Island, Antarctica, as a typical glacier in the south subpolar region. Salzburguer Geograph Mater 28: 73–80.

PATERSON WSB. 1994. The Physics of Glaciers. Oxford: Elsevier, 480 p.

PICHLMAIER M, AQUINO FE, SANTOS DA SILVA C AND BRAUN M. 2004. Suspended sediments in Admiralty Bay, King George Island (Antarctica). Pesq Antart Bras 4: 77–85.

SIMÕES JC, BREMER UF, AQUINO FE AND FERRON FA. 1999. Morphology and variations of glacial drainage basins in the King George Island ice field, Antarctica. Ann Glaciol 29: 220–224.

SIMÕES JC, FERRON FA, BERNARDO RT, ARISTARAIN AJ, STIÉVENARD M, POURCHET M AND DELMAS RJ. 2004. Ice core study from the King George Island, South Shetlands, Antarctica. Pesq Antart Bras 4: 9–23.