PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

Back-Arc Extension in the Southern Andes: A Review and Critical Reappraisal

I. W. D. Dalziel

Phil. Trans. R. Soc. Lond. A 1981 **300**, 319-335
doi: 10.1098/rsta.1981.0067

Phil. Trans. R. Soc. Lond. A **300**, 319–335 (1981)
Printed in Great Britain

Back-arc extension in the southern Andes: a review and critical reappraisal†

By I. W. D. Dalziel

Lamont-Doherty Geological Observatory of Columbia University,
Palisades, New York, NY 10964, *U.S.A.*

[Pullout 1]

The interpretation that the mafic 'rocas verdes' (green rocks) complex of the southern Andes represents part of the uplifted floor of a Late Jurassic to Early Cretaceous back-arc basin has proved particularly useful in understanding the geological evolution of the southern Andes, the north Scotia Ridge and the Antarctic Peninsula. Clear field evidence of the back-arc setting of the 'rocas verdes' gabbro-sheeted dyke – pillow lava ophiolitic assemblages has encouraged fruitful petrological and geochemical comparison with mid-ocean ridge and marginal basin basalts, other onshore ophiolite complexes, and Archaean greenstone belts.

Uncertainty still surrounds estimates of the original width and depth of the basin, as well as the proportion of new mafic crust, compared with relict sialic crust, in the basin floor. These questions are unresolved, owing mainly to the considerable Lower Cretaceous turbiditic basin infill and the effects of mid-Cretaceous compressional deformation.

While the field relations clearly indicate that the 'rocas verdes' basin is not an older piece of ocean floor 'trapped' behind a volcanic arc, it is not yet clear whether the basin is directly subduction-related or falls in the category of back-arc 'leaky transforms' like the proto-Gulf of California or apparent 'rip-off' features like the Andaman Sea.

1. Introduction

Gansser (1973) pointed out that the Andean Cordillera can be divided into three parts. In the central segment, the longest one, there are no Mesozoic or Cainozoic rocks with obvious oceanic characteristics, although some 'eugeosynclinal' assemblages are reported. In the northern and southern extremities of the Andes, however, rocks with oceanic affinities definitely do occur (figure 1). This paper is concerned with the Cordillera south of 50° S latitude where rocks originally mapped as the 'rocas verdes' (green rocks) have been interpreted as representing part of the floor of a Late Jurassic to Early Cretaceous back-arc basin analogous to those of the western Pacific today.

Although the geology of the depositional basins on the Atlantic side of the southern Andes has been reasonably well known for some time because of their hydrocarbon potential, it has been only since the mid-1960s that knowledge of the remote and inaccessible Cordillera has accumulated. Work along the Pacific margin of southernmost Chile (Katz & Watters 1966) led Katz (1973) to propose that the 'eugeosynclinal' rocks there (volcaniclastic turbidite, chert and mafic igneous rocks) might represent a terrain akin to the marginal basins of the western Pacific. Following the same line of thought, a party from Lamont–Doherty Geological

---

† Lamont-Doherty Geological Observatory Contribution no. 3084.

Observatory went to Chile during the 1972/3 austral summer with the object of studying the internal structure and field relations of the belt of 'rocas verdes' noted during reconnaissance mapping in Cordillera Sarmiento between 51° and 52° S latitude (figure 2, pullout 1). It was thus determined that the mafic zone consists of gabbro cut by a few mafic dykes passing upwards into a sheeted dyke complex. Above that there are dykes with screens of pillow lava grading upwards into a thick sequence of pillow lavas. Although ultramafic rocks are not exposed, it was felt that the internal structure of this complex justified its being referred to as the upper part of an ophiolite sequence. The main ophiolitic body appears to have sheared margins.

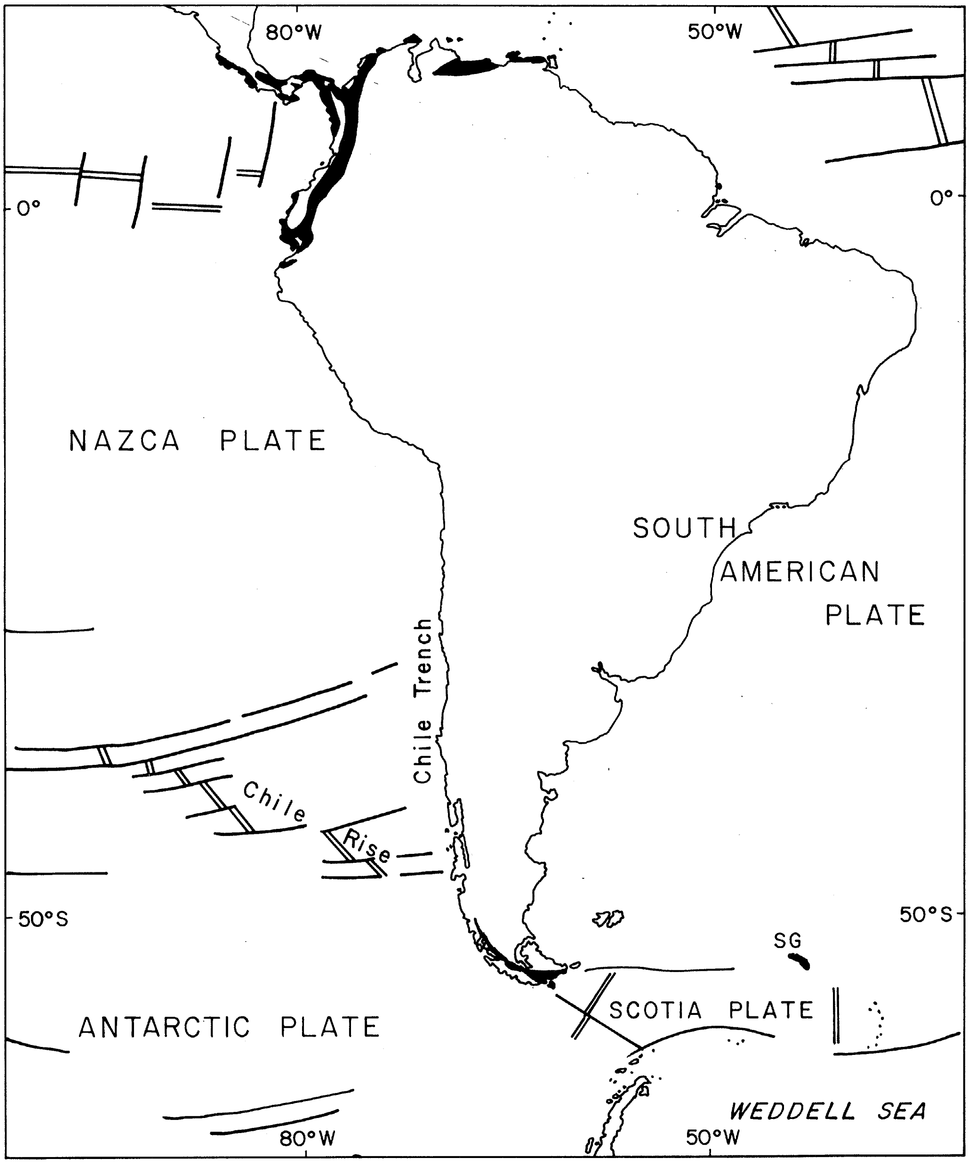


FIGURE 1. The location and present plate tectonic setting of Mesozoic and Cainozoic rocks of oceanic affinities in the Andes (modified after Gansser (1973)); SG, South Georgia; double lines, spreading centres; single lines, transform faults.

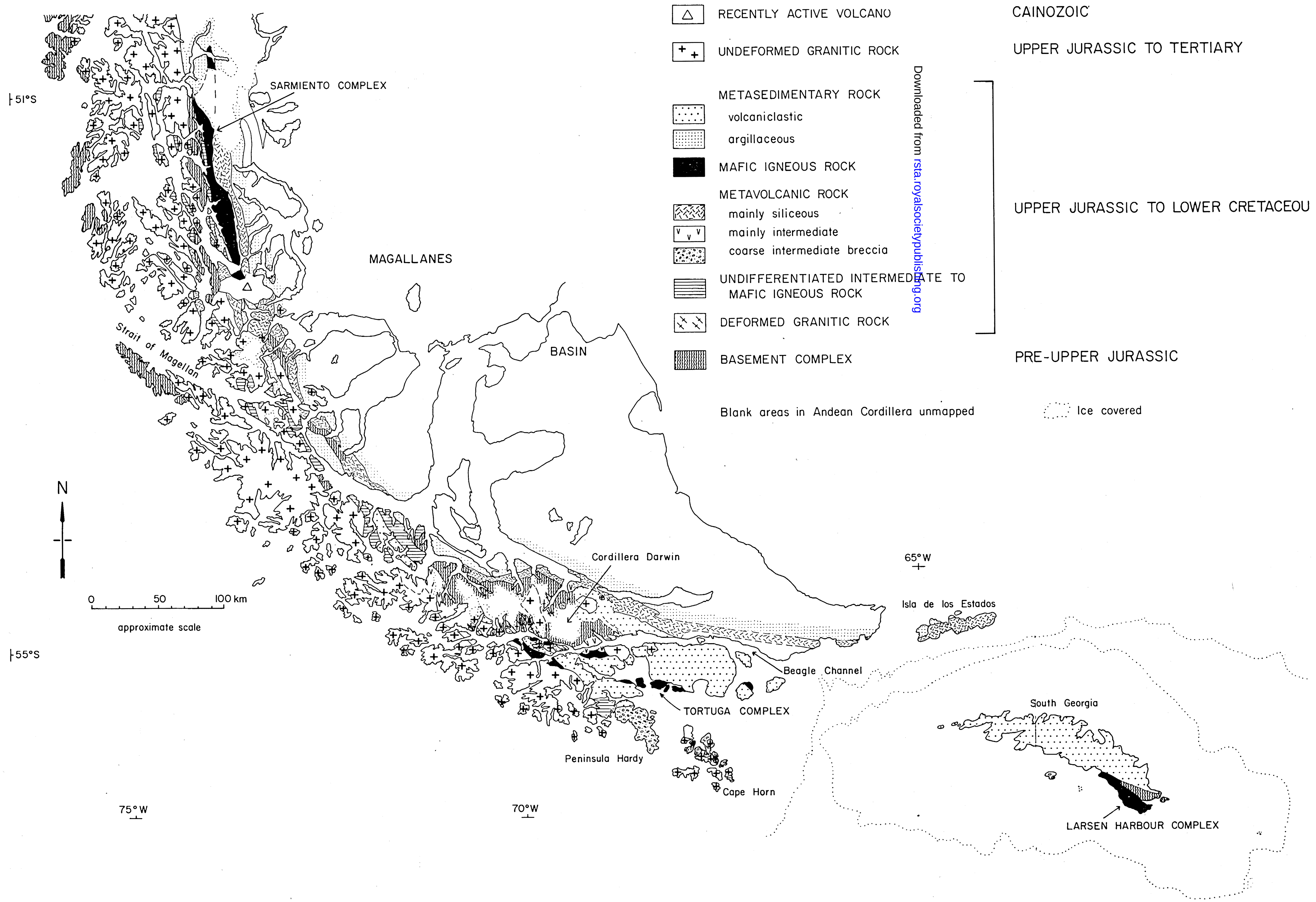
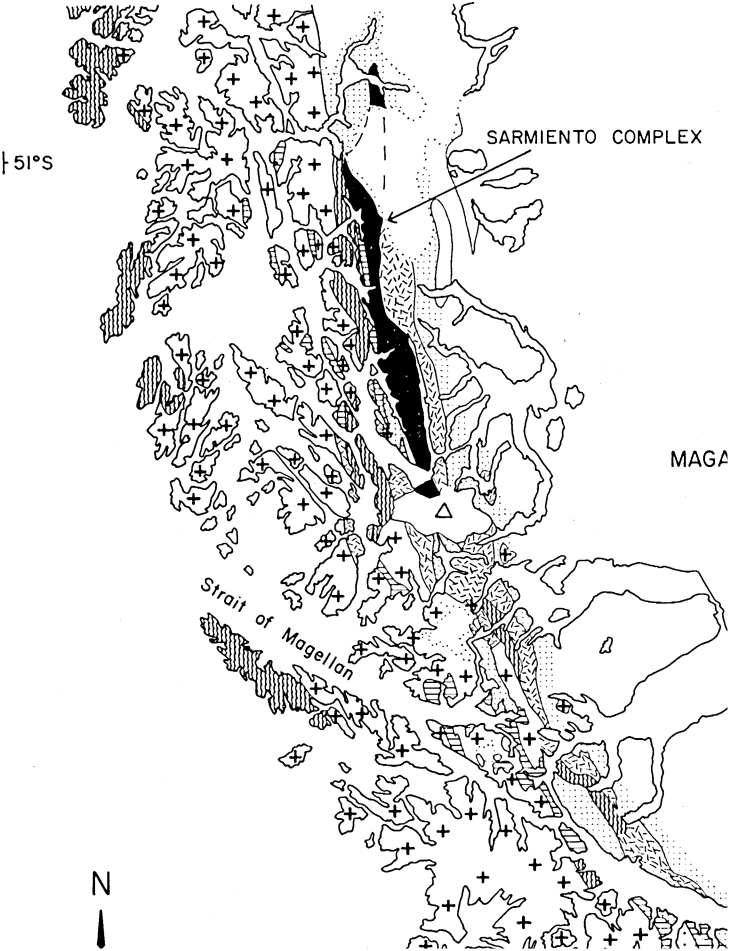
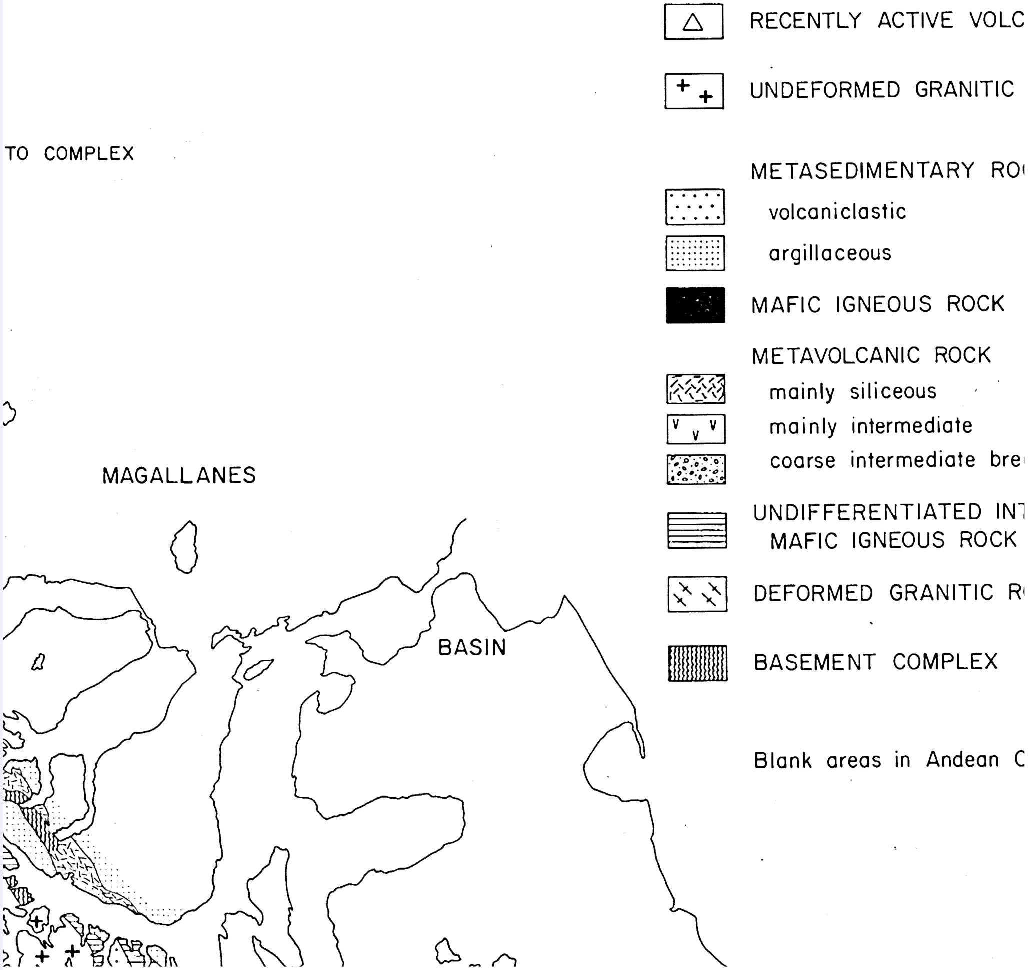


FIGURE 2. The geology of the southernmost Andes and South Georgia. The South Georgia platform is shown in its assumed approximate pre-drift position with respect to South America (see Dalziel *et al.* 1975; Tanner 1981).


51°S
SARMIENTO COMPLEX
MAGA
Strait of Magellan
N


TO COMPLEX
MAGALLANES
BASIN
RECENTLY ACTIVE VOLC
UNDEFORMED GRANITIC
METASEDIMENTARY RO
volcaniclastic
argillaceous
MAFIC IGNEOUS ROCK
METAVOLCANIC ROCK
mainly siliceous
mainly intermediate
coarse intermediate bre
UNDIFFERENTIATED IN
MAFIC IGNEOUS ROCK
DEFORMED GRANITIC R
BASEMENT COMPLEX
Blank areas in Andean C

ACTIVE VOLCANO

CAINOZOIC

IED GRANITIC ROCK

UPPER JURASSIC TO TERTIARY

MENTARY ROCK

astic

ous

NEOUS ROCK

CANIC ROCK

liceous

ntermediate

ntermediate breccia

ENTIATED INTERMEDIATE TO
GNEOUS ROCK

D GRANITIC ROCK

UPPER JURASSIC TO LOWER CRETACEOU

T COMPLEX

PRE-UPPER JURASSIC

as in Andean Cordillera unmapped

⬚ Ice covered

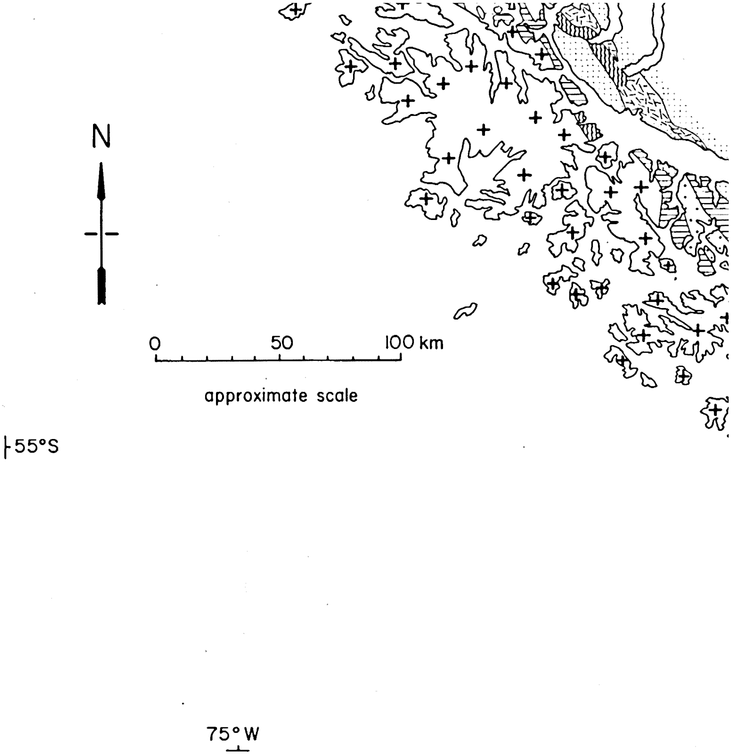


Figure 2. T
its assu
1981).

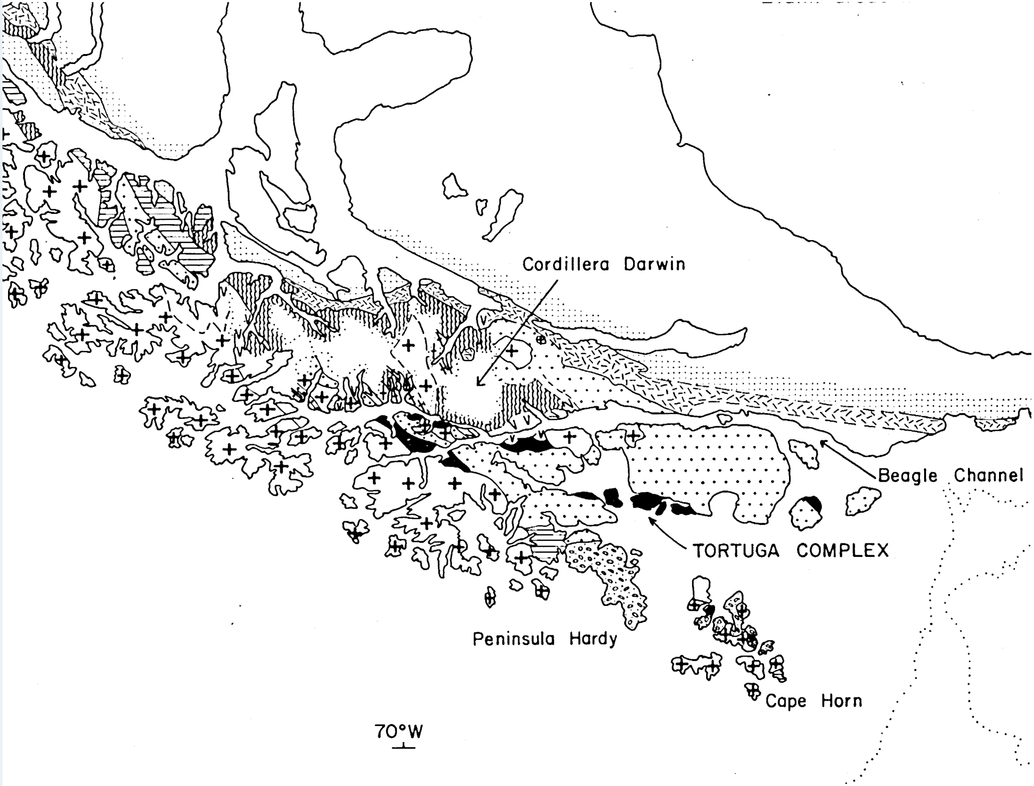
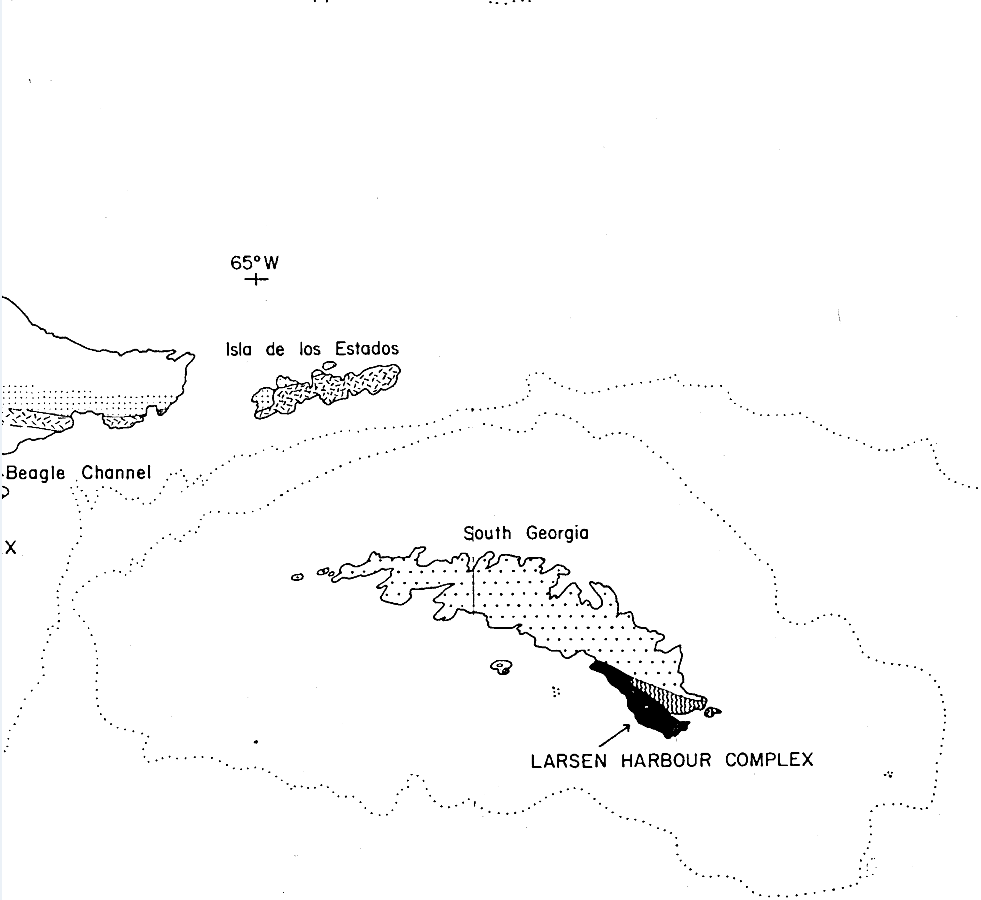


Figure 2. The geology of the southernmost Andes and South Georgia. The South Georgia platf
its assumed approximate pre-drift position with respect to South America (see Dalziel *et al.*
1981).



n Georgia platform is shown in
ee Dalziel *et al.* 1975; Tanner

Fringeing mafic masses, however, clearly intrude silicic volcanic rocks of probable Late Jurassic age (the Tobifera Formation) on the continental side of the upper Mesozoic Patagonian batholith. The volcanics unconformably overlie a previously deformed and metamorphosed basement. These field relations led to the ophiolitic complex being interpreted as the floor of a marginal basin in which the Lower Cretaceous volcaniclastic turbidites and minor chert of Katz's 'eugeosyncline' were deposited (Dalziel *et al.* 1974). In order to make the point that the basement intruded by the mafic rocks was an integral part of the South American continent rather than part of some exotic terrain at the time that the basin formed, the Japanese volcanic arc and the Japan Sea were suggested as a modern analogue for the Late Jurassic to Early Cretaceous island arc and marginal basin in the southernmost Andes.

This concept has proved remarkably helpful in the elucidation of the regional geology of the southern Andean Cordillera (Bruhn & Dalziel 1977; Bruhn 1979; Dalziel & Palmer 1979; Suárez 1979; Andrews-Speed 1980), the northern limb of the Scotia Ridge (Dalziel *et al.* 1975; Suárez & Pettigrew 1976; Bell *et al.* 1977; Storey *et al.* 1977; Winn 1978; Tanner & Rex 1979; Tanner 1981; Tanner *et al.* 1980), and the Antarctic Peninsula (Dalziel 1974; Suárez 1976). Moreover, the concept has stimulated petrological and geochemical comparison of the southern Chilean ophiolitic rocks, not only with other onshore ophiolites, but also with mid-ocean ridge basalts (de Wit & Stern 1978; Elthon 1979; Saunders *et al.* 1979; Elthon & Ridley 1980), continental basalts (Bruhn *et al.* 1978), and Archaean greenstone belts (Tarney *et al.* 1976). It has led to studies of ocean floor metamorphism that have stimulated ideas on hydrothermal circulation in oceanic crust (Elthon & Stern 1978; Stern & Elthon 1979; Elthon 1980) and the origin of seismic layering and magnetic anomalies (de Wit & Stern 1976; Stern *et al.* 1976). Finally, as will be discussed here, it is helping to shed some light on the causes of marginal basin development and destruction (Dalziel *et al.* 1974; Dewey 1981).

The contributions concerning this remote but important back-arc basin are widely scattered in the literature, and some of the conclusions and interpretations are rather divergent. It seems worth while in this volume to review the field evidence and indulge in a critical re-appraisal of the basic conclusions reached with regard to its significance. The uncertainties seem to revolve around the original dimensions of the basin and the proportion of new mafic as opposed to relict sialic rocks flooring it, rather than around its very existence or its back-arc setting. Hence the concluding discussion addresses the possible causes of Late Jurassic to Early Cretaceous back-arc spreading in the southern Andes within a global framework.

2. TIME–SPACE SETTING OF THE BACK-ARC BASIN

The present extent of rocks associated with the back-arc basin in the southern Andes is from 50° S latitude to Cape Horn at 56° S (figure 2). After the Patagonian orocline has been taken into account, this amounts to over 700 km. An additional 50 km or more may lie on the continental shelf of South America east of Cape Horn and 350 km on the partly submerged South Georgia platform (Simpson & Griffiths 1981). South Georgia is currently situated approximately 1700 km east of Cape Horn (figure 1) but is widely accepted to be a former continuation of the back-arc terrain in South America (figure 2) (Dalziel *et al.* 1975; Suárez & Pettigrew 1976; Bell *et al.* 1977; Storey *et al.* 1977; Tanner & Rex 1979; Tanner 1981), moved into its present position by transform motion along the North Scotia Ridge (figure 1).

The evidence suggests that the basin opened before, but probably close to, the Jurassic–

Cretaceous boundary (140 Ma), that it existed as a site of volcaniclastic turbidite accumulation until the late Early Cretaceous (110–100 Ma), and that it was being uplifted and destroyed by mid-Cretaceous times (100–90 Ma). This means that it developed along the southernmost part of the South American segment of Gondwanaland about the time that western Gondwanaland (South America–Africa) broke away from eastern Gondwanaland (India–Australia–East Antarctica) and before the South Atlantic started to open. It ceased to exist at about the time of the change in the pole of rotation of South America with respect to Africa (Rabinowitz & LaBrecque 1979) and the time of the worldwide increase in sea floor spreading rates (Larson & Pitman 1972).

Since the Nazca – South America – Antarctica triple junction at the Chile Rise – Chile Trench intersection (figure 1) must have moved along the Pacific boundary of the South American plate over the past 20–30 Ma (Herron & Tucholke 1976), it must be assumed that the plate configuration in the southeastern Pacific Ocean basin at the time of opening of the basin was basically comparable to that of the central Andes of today with the ancestral Nazca (Farallon) plate being consumed beneath the South American plate. This does not, however, preclude the existence of other elements such as ‘leaky-transform’ faults or spreading ridges that have since been subducted but that might have influenced the tectonics of the South American margin at the time. Also, while the origin of Patagonian orocline is still uncertain, it now seems likely that there was a major bend in the South American continental margin there even before the break-up of the Gondwana continent (Dalziel 1981).

Mention should also be made of the history of the region before and after the development of the back-arc basin in the Late Jurassic to Early Cretaceous. The profound unconformity beneath the Middle to Upper Jurassic volcanic rocks that are cut by some of the mafic intrusions associated with the development of the basin indicates a widespread and important tectonic event of unknown nature in the history of Gondwanaland. Perhaps it resulted from uplift related to the incipient break-up of the supercontinent. Whatever the cause, it seems to have marked merely one event in a complex but essentially continuous history of subduction going back into at least the mid-Palaeozoic. The pre-Middle Jurassic basement of southern South America clearly records convergence with its accreted fore-arc terrain as well as calc-alkaline igneous rocks (Barker *et al.* 1976; Dalziel & Forsythe 1977; de Wit 1977; Forsythe & Mpodozis 1979; Dalziel 1981). Likewise the presence of numerous post-Middle Cretaceous calc-alkaline plutons throughout the southern Andes, including some within the uplifted back-arc basin terrain, testify to subduction having continued beneath the South American plate after the basin ceased to exist (Tanner & Rex 1979; Nelson *et al.* 1980). The present situation seems to involve very slow subduction south of the Nacza – South America – Antarctica triple junction (Forsyth 1975).

The initiation of a back-arc basin in the southernmost Andes in the latest Jurassic seems to have been a unique event in this long history of subduction. No comparable terrain has been recognized in the basement rocks. It is interesting to note that the development of the marginal basin coincides with the ages of the Karroo and Ferrar mafic igneous activity and with the incipient break-up of Gondwanaland. Hence all these events may be related to each other (Cox 1978; Dalziel 1981).

3. Critical field relations

The main mafic bodies in southern South America known as the Sarmiento and Tortuga complexes (figure 2) have now been mapped on a scale of 1:100000 and 1:250000 respectively (Allen 1980; Elthon & Ridley 1980). Between them there is a zone 400 km long that is less well known, but it is unlikely that any other ophiolitic segments of comparable size exist there. British Antarctic Survey geologists have completed mapping a third major body, the Larsen Harbour Formation, on South Georgia Island (Bell *et al.* 1977; Storey *et al.* 1977; Anon. 1980; Tanner *et al.* 1980).

The geological framework of the South American segment of Gondwanaland into which the Sarmiento, Tortuga and Larsen Harbour complexes were emplaced consisted of a basement containing rocks as young as Permian unconformably overlain by a layer of volcanic rocks interbedded with shallow marine sedimentary layers containing Kimmeridgian to Tithonian fossils (Natland *et al.* 1974). As previously mentioned, the basement seems to comprise the fore-arc terrain of a pre-Middle Jurassic convergent plate margin along the Pacific margin of Gondwanaland. It consists of 'slices' of mafic pillow lavas and bedded chert, together with fusulinid limestone, tectonically intercalated within a greywacke–shale terrain. The volcanic cover is dominantly silicic in composition, but a narrow zone of lavas and pyroclastic rocks of intermediate composition crops out within the present-day Cordillera (Bruhn *et al.* 1978; Nelson *et al.* 1980). In addition, there are granitic rocks apparently pre-dating emplacement of the ophiolitic complexes on the continental side of the back-arc basin terrain in Tierra del Fuego and of the Larsen Harbour complex on South Georgia (Dalziel & Cortés 1972; Tanner & Rex 1979; Nelson *et al.* 1980). It seems likely that these granitic rocks are co-magmatic with the Upper Jurassic volcanic rocks and represent subvolcanic plutons. The high initial ratio of $^{87}Sr/^{86}Sr$ of the granitic rocks in Tierra del Fuego supports this interpretation (Hervé *et al.* 1979). There is also a chemical distinction between arc-related calc-alkaline rocks and back-arc basin related basaltic material (Bruhn *et al.* 1978).

It should be re-emphasized that the main mafic bodies of the Sarmiento, Tortuga and Larsen Harbour complexes mainly have tectonized contacts, The Sarmiento complex is in fault contact with the basement rocks and their volcanic cover, the Tortuga complex has been upfaulted into Lower Cretaceous volcaniclastic turbidites, the Larsen Harbour rocks also have a faulted contact along their northeastern side, but they may be transitional downwards into the island arc rocks (Annenkov Island Formation) to the southwest (Tanner *et al.* 1980). Mafic extrusives are interbedded with the Upper Jurassic silicic to intermediate volcanic rocks. On the other hand, mafic volcanic and intrusives within the Lower Cretaceous turbidites are scarce. Hence it appears that the ophiolitic rocks were emplaced during, perhaps late in, the history of Late Jurassic silicic and intermediate volcanism and covered by the Lower Cretaceous turbidites. These latter sedimentary rocks contain andesitic detritus testifying to arc volcanism's having continued after development of the marginal basin near the Jurassic–Cretaceous boundary (Dalziel *et al.* 1975). The turbidites rest in places directly upon lavas of the ophiolitic sequence, for instance in the southern part of the Sarmiento complex and in reefs off the southwestern coast of South Georgia (R. Allen and P. W. G. Tanner, personal communications, 1980). Chert is not extensively developed, but occurs locally within the pillow lavas, on top of them, and within the turbiditic layers.

Hence it seems reasonable to interpret the mafic rocks of the southern Andes as having been

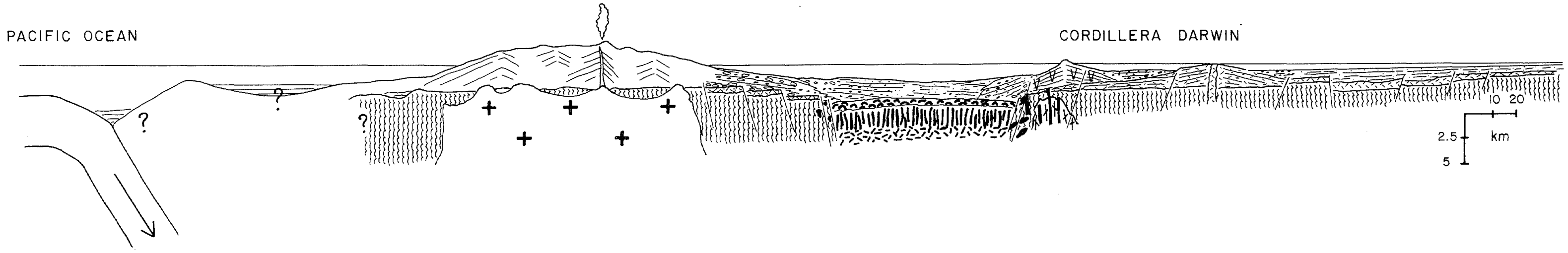


FIGURE 3. Interpretive cross section of the southernmost Andes and South Georgia in the Early Cretaceous; diagram shows the maximum possible width of mafic crust based on *present-day* outcrop (for ornament see figure 2).

emplaced within a subduction-related volcanic arc along the Pacific margin of the South American segment of Gondwanaland (figure 3). Whether or not an actual 'remnant arc' existed in the topographic sense that is implied by Karig (1971) for the western Pacific marginal basins, there seems to be no doubt that Upper Jurassic volcanics, and possibly subvolcanic plutons, were separated by the emplacement of the mafic igneous rocks from the site of arc volcanism during the Early Cretaceous. In this sense at least, a remnant arc seems to have existed in the southern Andes and South Georgia. Indeed, the back-arc basin seems to have opened along the Pacific edge of the Late Jurassic magmatic arc.

It could be argued that the mafic igneous bodies of the southern Andes and South Georgia should not strictly be termed ophiolites since ultramafic rocks are absent. Nonetheless, it seems reasonable to refer to them as the *upper part* of ophiolitic bodies since only the top few hundred metres of gabbro are seen beneath sheeted dykes, pillow lavas and volcaniclastic turbidites, and the geochemistry of the extrusive mafic rocks requires the presence of ultramafic cumulates at depth. Moreoever, the dykes are vertical and for the most part trend parallel to the Cordillera; hence the bodies appear to have been uplifted vertically and it seems likely that the apparent absence of ultramafic rocks is merely a function of exposure level. Nonetheless it is possible that the exposed mafic bodies are connected only by feeder dykes through the continental basement to their ultramafic parent. It may well be that the Sarmiento, Tortuga and Larsen Harbour complexes represent stages in a transition from the mafic igneous rocks of the central Andean 'eugeosynclines' to the floor of a real marginal ocean basin.

4. Dimensions of the basin and nature of its floor

The present width of the terrain across which ophiolitic rocks crop out varies from a minimum of 5 km in the northern part of the Sarmiento Complex to a maximum of 50 km between the Beagle Channel and the Hardy Peninsula on Isla Hoste in the south (figure 2). After the apparent virgation in the latter area has been taken into account (figure 2), the width may reach 100 km before the terrain is truncated by the South American shelf edge. Both geological and geophysical data indicate, however, that the arc bounding the back-arc basin on its Pacific side is present on the South Georgia platform (Suárez & Pettigrew 1976; Tanner 1981; Simpson & Griffiths 1981; Tanner *et al.* 1980). Hence the widening towards the south and east seen on the South American mainland may not continue as far as the original position of the South Georgia platform. Any former continuation of the basin beyond this platform is obscure. De Wit (1977) suggested that the central part of the floor of the present Scotia Sea between the South Georgia and South Orkney Island platforms represents a continuation of the marginal basin floor. Recent identification of marine magnetic anomalies indicates, however, that the crust of the central Scotia Sea is Cainozoic (Barker & Hill, this symposium). One possibility is that in some way or another the back-arc basin in South America connected with a similar feature in the western Weddell Sea before the Cainozoic development of Drake Passage and the Scotia Sea (Dalziel 1974; Suárez 1976). It is at present impossible to substantiate this suggestion. The connection, if it existed, seems likely to have been indirect. A re-entrant in the Pacific edge of the Gondwana shield in the vicinity of the Cainozoic Scotia arc (Dalziel 1981) suggests that the connection between the Antarctic Peninsula and southern South America may once have been similar to that between the Alaskan–Aleutian and Kamchatkan–Kurile arc systems.

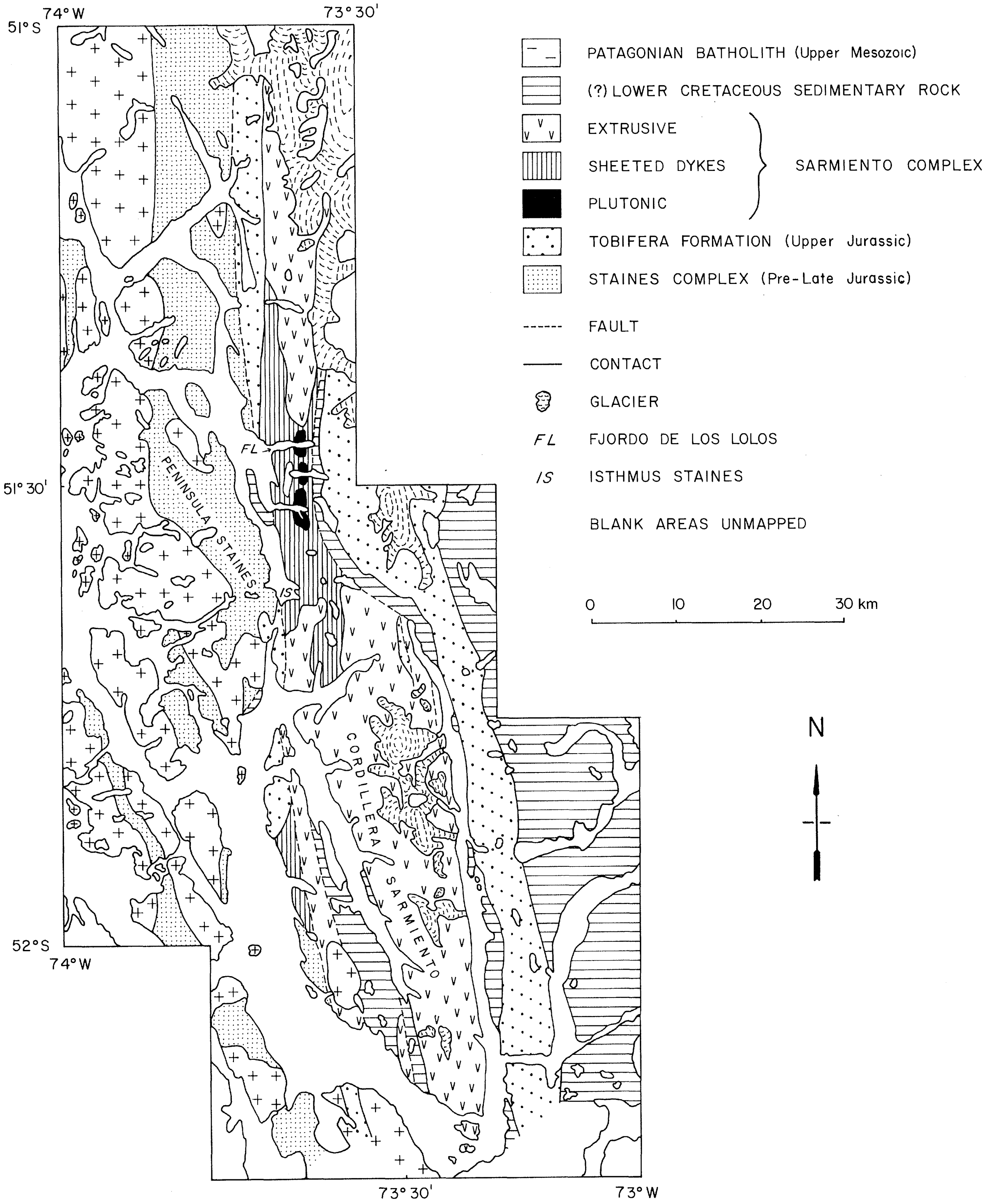


Figure 4. Geological map of the Sarmiento complex and surrounding area (after Allen 1980).

The present width of the ophiolitic zone must be considerably less than the width of the original back-arc basin; first, because the sedimentary infilling of the basin extended onto the rifted margin of the South American mainland (Bruhn 1979; Nelson *et al.* 1980) (figures 2 and 3), and secondly because the mid-Cretaceous deformation certainly resulted in tectonic shortening of the rocks perpendicular to the continental margin and may have involved décollement and subduction, or at least underthrusting, of part of the basin floor.

The width of the main ophiolitic body (gabbro and sheeted dykes) in the northern part of the Sarmiento complex appears to be no more than 5 km (figure 4). As previously mentioned, the margins are sheared. Deformation of the rocks was concentrated in certain zones but consisted mainly of east–west flattening (R. Allen, personal communication, 1980). Even allowing for extreme tectonic shortening, it seems rather unlikely that the width of mafic rocks forming the basin floor here was ever more than a few kilometres unless some of them have been underthrust beneath the arc terrain to the west or the older volcanic ('remnant arc') terrain to the east.

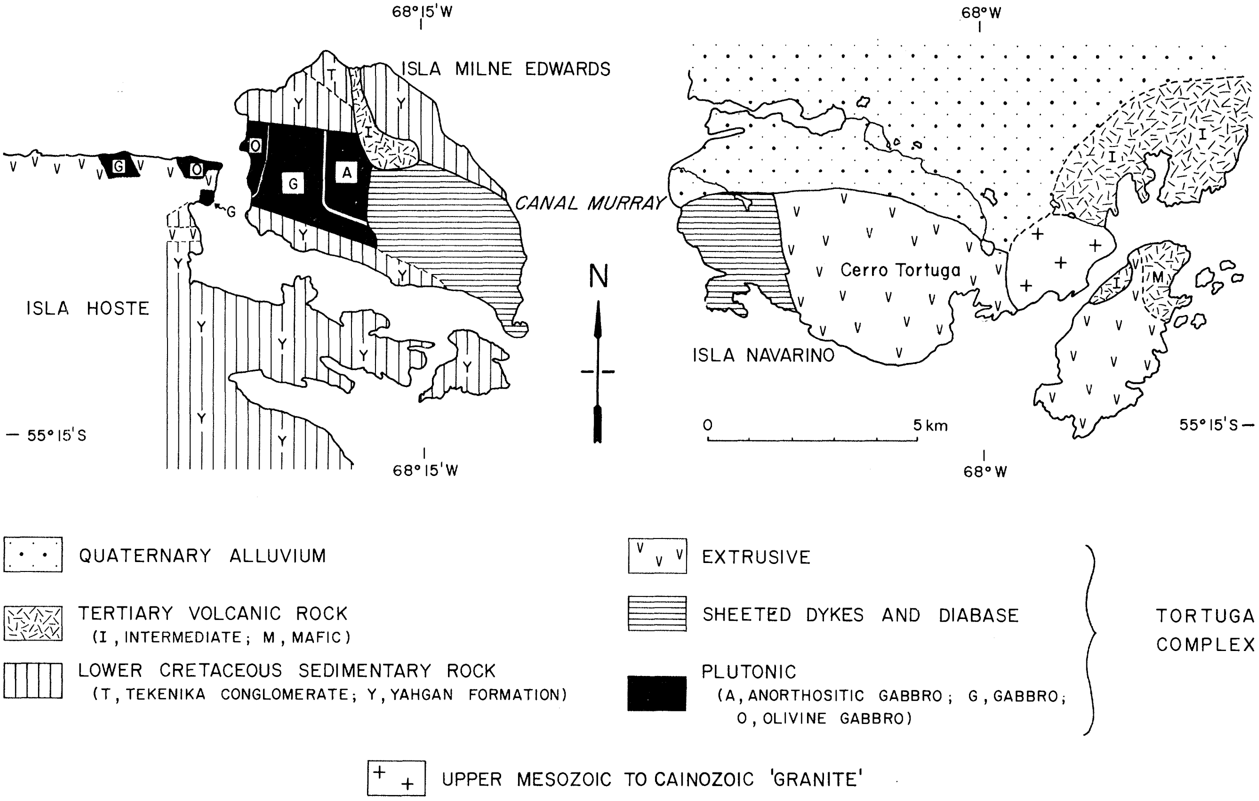


Figure 5. Geological map of the Tortuga complex and surrounding area (after Elthon & Ridley 1980).

Although the width of the terrain with ophiolitic outcrops in southern Tierra del Fuego is 50 km (figure 2), Lower Cretaceous turbidites crop out across most of this zone. The exposures of intrusive mafic rock upfaulted into the sedimentary cover amount to only 5 km (figure 5). Hence it is not possible to say exactly what proportion of the floor of this part of the sedimentary basin is in fact ophiolitic (Andrews-Speed 1980). Certainly the turbidites north of the Beagle Channel rest on the pre-Middle Jurassic continental basement across a present-day outcrop width of up to 50 km (Nelson *et al.* 1980) (see figure 3). The Larsen Harbour complex on South Georgia has a maximum width of 10–15 km including its possible submarine extension (figure 6).

In Tierra del Fuego and on South Georgia the turbiditic infilling of the basin is intensely

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

folded with a vergence dominantly away from the Pacific (Dalziel *et al.* 1975; Tanner 1981). Structures in the highly deformed 'remnant arc' terrain along the continental edge of the back-arc basin also have a vergence dominantly towards the Atlantic (Bruhn & Dalziel 1977; Bruhn 1979; Dalziel & Palmer 1979; Nelson *et al.* 1980). Hence it could be suggested that part of the floor of a back-arc basin considerably wider than the present outcrop zone of ophiolitic rocks was subducted beneath the arc on the Pacific side. At present it appears likely that in South America the rocks of the basin floor and infill have primarily been shortened and uplifted towards the continent along a steep zone of reverse faulting coinciding with what must have been the zone of most concentrated normal downfaulting during the inception of the basin (Bruhn & Dalziel 1977; Bruhn 1979; Dalziel & Palmer 1979). Field relations on South Georgia, where the basin may have been wider, are more suggestive of subduction (Tanner *et al.* 1980).

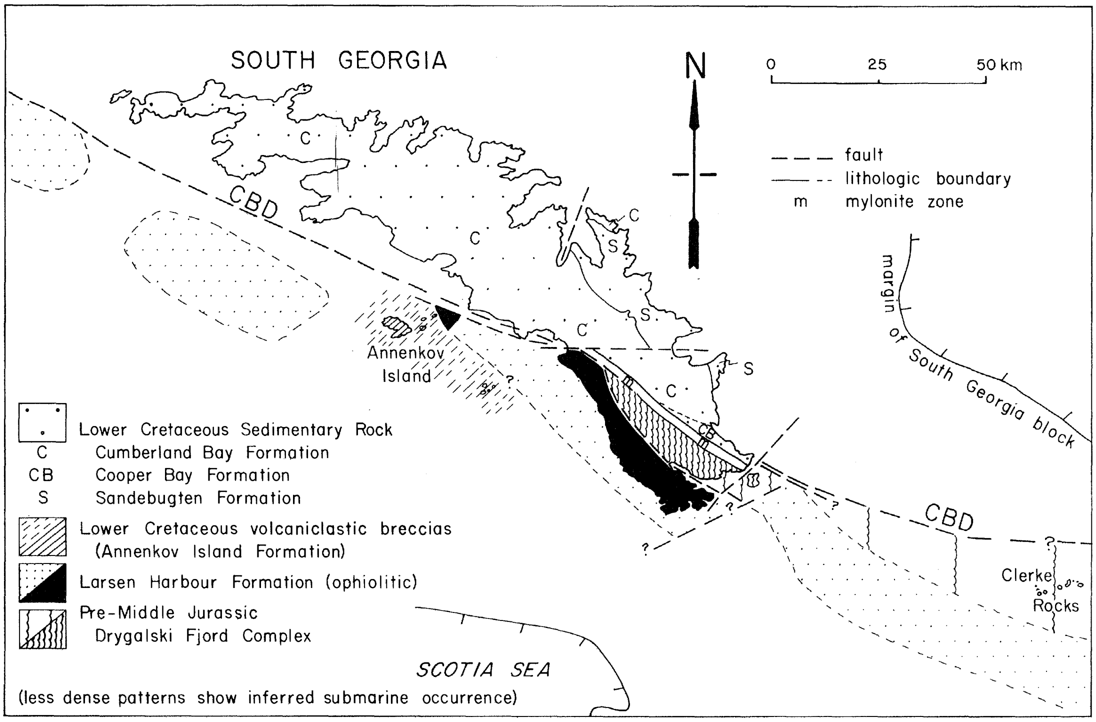


FIGURE 6. Geological map of the South Georgia continental block (after Tanner *et al.* 1980); CBD, Cooper Bay dislocation.

The intensity of the folding, cleavage development and local thrusting in the turbiditic basin infill in Tierra del Fuego and on South Georgia indicates that considerable shortening of the sedimentary pile must have occurred. Detailed structural studies on South Georgia may give a more quantitative estimate; meanwhile one might suggest that about 50% shortening may have been involved. There is a problem in reconciling shortening of this order of magnitude in the sedimentary cover with the absence of penetrative deformation in the ophiolitic Tortuga complex on Isla Navarino. Reconciliation of this enigma could be in part explained by décollement and underthrusting (i.e. subduction) of part of the basin floor, or else by décollement and concentration of basement shortening in narrow but intense shear zones. The latter explanation does have some support in the observation of shear zones locally (Bruhn 1979), the former by the substantial shortening of the turbidites on South Georgia (Dalziel *et al.* 1975; Tanner 1981; Tanner *et al.* 1980).

Uncertainties also surround the question of the original depth of the basin. Comparison with present-day back-arc basins around the world suggest a minimum water depth to a mafic floor of 1.0–1.5 km. Trace fossils in the sedimentary infill of the basin indicate a bathyal depth (Winn 1978). Estimates of the thickness of the turbidites vary. In the southern part of the Sarmiento complex where the turbidites overlie mafic pillow lavas, the turbidites are probably a minimum of 1 km thick with no top exposed (R. B. Allen, personal communication, 1980). There are several kilometres present on South Georgia (Dalziel *et al.* 1975; Tanner 1981), and even though this is likely to be a prograded sequence the thickness seems to support something considerably greater than continental shelf depth for the back-arc basin.

In summary, the proportion of new mafic (i.e. ophiolitic) compared with relict South American continental crust forming the floor of the back-arc basin beneath the present zone of outcrop of ophiolitic rocks is uncertain owing to the extent of the turbiditic infill. Only about 10% of the width is demonstrably mafic. The original width of the basin is hard to estimate owing to the subsequent compressional deformation. It is likely to have been considerably wider than the present exposure width of the associated rocks, perhaps twice as wide. There is no independent evidence, however, to indicate that hundreds of kilometres of mafic back-arc basin crust were lost by subduction. The depth of the basin is likely to have been comparable to that of present day back-arc marginal seas. Hence the overall dimensions of the Early Cretaceous back-arc basin in the southern Andes may perhaps have been comparable to those of the Andaman Sea or Bransfield Strait rather than the Sea of Japan, but a larger basin is not ruled out.

5. Destruction of the basin

The youngest rocks of the marginal basin infilling are upper Lower Cretaceous or lower Upper Cretaceous (Dott *et al.* 1977). Together with fine-grained Albian sedimentary strata on the continental side of the marginal basin, these were deformed before the intrusion of 80–90 Ma granitic plutons. The first appearance of flysch-type strata on the Atlantic side of the Cordillera occurred in the Cenomanian (Scott 1966), indicating that this was the time of the initial deformation and tectonic uplift of the arc, marginal basin, and 'remnant arc' terrains (Dalziel & Palmer 1979). Before this mid-Cretaceous tectonism, only extensional deformation had affected southern South America since the extrusion of the Upper Jurassic volcanics. Hence this marks a major change.

The nature of the mid-Cretaceous compressional deformation has been described in a number of recent papers (Dalziel *et al.* 1975; Bruhn & Dalziel 1977; Bruhn 1979; Dalziel & Palmer 1979; Nelson *et al.* 1980; Tanner 1981). Briefly, it appears that vertical uplift, strike-slip displacements and shortening at right angles to the Pacific margin were all involved. The arc, back-arc basin and 'remnant arc' terrains were all uplifted relative to the stable interior of the South American continent. Between the uplifted Cordillera and the stable interior a classical 'foredeep' developed known as the Magallanes basin (Natland *et al.* 1974). This foredeep was the site of deposition of flysch-like submarine fan deposits starting in the Cenomanian (Natland *et al.* 1974; Scott 1966; Winn 1978). While there is evidence of the uplift of the arc terrain with respect to the marginal basin terrain on South Georgia (P. W. G. Tanner, personal communication, 1980), in general differential uplift between the island arc and back-arc basin terrains does not seem to have been extensive because coarse proximal volcanic breccias are still preserved along the rear (continental) edge of the arc (Suárez & Pettigrew 1976; Dott *et al.* 1977; Suárez 1979). As previously mentioned, however, it appears that the arc and back-arc basin terrains were uplifted significantly (perhaps in excess of 5 km) with respect to the continental edge of the basin (?remnant arc terrain). Finally, local intensification of folding and cleavage development, accompanied by deviations in strike associated with major features such as the lineament along the western end of the Strait of Magellan, suggest that strike-slip displacements may also have been important at that time.

For the most part the structures developed in the mid-Cretaceous in the rocks of the arc and back-arc basin terrains are comparatively simple. Local complications are often associated with shear zones. Volcanics in the arc terrain are gently folded and sometimes have a well developed slaty cleavage trending parallel to the continental margin and dipping steeply (Dalziel & Cortés 1972, plate 2B). The turbiditic infill of the marginal basin is dominated by tight asymmetric folds with hinge lines parallel to the continental margin and verging toward the continent (Dalziel *et al.* 1975; Tanner 1981). The Sandebugten sedimentary sequence on South Georgia Island appears to have a strong vergence towards the Pacific (Dalziel *et al.* 1975; Tanner 1981), but this may be due to subsequent refolding (P. W. G. Tanner, personal communication, 1980). The Sandebugten rocks contain siliceous (rather than andesitic) volcanic detritus and were apparently derived from the continental interior (Dalziel *et al.* 1975; Winn 1978). The affinities of the unfossiliferous Sandebugten rocks are still uncertain, however, and the possibility still exists that they are part of the pre-Jurassic basement terrain.

Mid-Cretaceous structures in the highly deformed 'remnant arc' terrain are complex and polyphase (Bruhn & Dalziel 1977; Bruhn 1979; Dalziel & Palmer 1979; Nelson *et al.* 1980) Intense cleavage and associated folding verge strongly towards the continent, consistent with the apparent uplift of the back-arc basin terrain with respect to the 'remnant arc'. Subsequent structures either reinforce this vergence (Dalziel & Palmer 1979) or else are conjugate, involving 'backfolding' and 'backthrusting' in Alpine terms (Nelson *et al.* 1980). Nonetheless, the parallelsim of fold hinge lines and bedding–cleavage intersections suggests that these structural 'phases' represent part of a sequence of progressive strain rather than being the result of independent deformational pulses.

On the continental side of the back-arc basin terrain in Cordillera Darwin, southern Tierra del Fuego, there is a culmination in the pre-Middle Jurassic basement that resulted from such polyphase, ?progressive deformation, including 'backfolding' towards the Pacific (Nelson *et al.* 1980). This culmination is of particular interest because it is the site of the highest-grade metamorphism in the southern Andes. Almandine–amphibolite facies rocks with staurolite and kyanite occur here together with migmatitic zones. The culmination is broadly similar to the 'metamorphic core complexes' of the North American Cordillera (Davis & Coney 1979), in other words a complex of high-grade metamorphic rocks localized in the 'core' of the orogen.

The exact significance of the high-grade metamorphic rocks in Cordillera Darwin has yet to be determined. It is of interest here, however, in view of the speculation concerning the North American complexes, to note that Cordillera Darwin is located on the continental edge of the back-arc basin terrain, and to note also that the intense compressive deformation, as well as the high-grade metamorphism, occurred after the extensional tectonics associated with back-arc basin development.

After the mid-Cretaceous deformation in the Andean Cordillera and the temporally and spatially related downwarp of a foredeep, the sediments filling the foredeep were deformed in the early and mid-Cainozoic. This resulted in a classic foreland fold and thrust belt comparable to that of the Canadian Rockies (Winslow 1981). The same problems with regard to extensive shortening of the cover and yet apparent lack of basement involvement arise in both places. It is possible that Pacificward underthrusting of continental basement beneath the Cordillera (so-called A-subduction) may account for the apparent enigma in the southern Andes. Alternatively, Winslow (1980) has recently suggested compressional reactivation of the Late Jurassic to Early Cretaceous normal faults associated with back-arc basin development. While this certainly appears to have taken place along the inner (continental) wall of the back-arc basin during the mid-Cretaceous deformation (Bruhn & Dalziel 1977; Bruhn 1979; Dalziel & Palmer 1979; Nelson *et al.* 1980), there is less evidence in support of such a mechanism in the Cainozoic.

6. Conclusions and discussion

From this reappraisal of the evidence it seems clear that an extensional basin, floored at least in part by new mafic crust, opened within a calc-alkaline volcanic arc along the Pacific margin of the South American segment of Gondwanaland during the latest Jurassic. It is not certain how wide the basin was, how deep it was, how far it extended towards the south and east, and what proportion of its floor consisted of new mafic crust as opposed to relict South American sialic crust. In a morphological sense at least, the most realistic modern analogues are the Andaman Sea and the Bransfield Strait. The exposures of new crust emplaced during the formation of the basin consist mainly of basaltic pillow lava, diabase dykes and gabbro. The absence of ultramafic rocks can be explained by the exposure level, and it seems reasonable to refer to the exposed mafic complexes as the upper part of ophiolitic bodies. The absence of ultramafics in no way invalidates the conclusion that a basin opened within the continental margin arc and existed as a back-arc basin along the Pacific margin during much of the Early Cretaceous. The basin was deep enough to accumulate a thickness of several kilometres of volcaniclastic turbidites derived principally from the island arc on its Pacific side. It may constitute a transition from the 'eugeosynclinal' basins of the central Andes to a truly oceanic marginal basin.

The basin was destroyed during the mid-Cretaceous compression of the arc, back-arc and passive continental margin (?remnant arc) terrains involving uplift of the arc and back-arc basin with respect to the 'remnant arc' and interior of the South American continent. It is uncertain whether underthrusting (subduction) of part of the basin floor took place in South America. The available evidence seems to require it only in South Georgia. Very likely the mechanism of basin closure was transitional from the narrow northern segment to the apparently wider southeastern segment on South Georgia.

There has been much discussion in the literature over the origin of back-arc basins (Scholz *et al.* 1971; Wilson & Burke 1972; Uyeda & Miyashiro 1974; Molnar & Atwater 1978; Chase 1978; Uyeda & Kanamori 1979; Jurdy 1979; Dewey & Windley 1981). Ignoring those formed by 'trapping' of a pre-existing segment of oceanic lithosphere, the development of a back-arc basin seems to require an extensional stress régime within and/or behind the volcanic arc on the upper plate at a convergent boundary. There seem to be many ways of arriving at this situation, and of changing it so that the back-arc spreading stops or even that the basin is compressed, uplifted and hence destroyed as happened in the southern Andes.

Confining the discussion to subduction-related back-arc basins, one can list many factors that will influence the state of stress in the upper plate (table 1). These factors are in turn controlled by four types of parameter: (1) relative plate motion; (2) the nature and configuration of the lower plate; (3) the nature and configuration of the upper plate; and (4) so-called 'absolute' plate motion. With regard to 'absolute' plate motion it should be borne in mind that the Late Jurassic to Early Cretaceous back-arc basin in the southern Andes is one of only four well documented back-arc basins associated with west-facing arc systems, the others being the Andaman Sea, Bransfield Strait and the Gulf of California, or perhaps more accurately the proto-Gulf. Hence there may be some background influences, such as the effect of Earth rotation on asthenospheric flow, that can locally be enhanced, or else must locally be overcome, to allow back-arc basin formation. The factors in table 1 are thus not intended to be regarded as being listed in order of their presumed importance.

TABLE 1. FACTORS INFLUENCING THE STATE OF STRESS IN THE UPPER PLATE AT A CONVERGENT BOUNDARY

influencing factor	*basin formation* (*extension*)	*basin inactive* (*neutral*)	*basin destruction* (*compression*)
(*a*) convergence rate	low	⟷	high
(*b*) convergence direction	oblique	⟷	direct
(*c*) age of subducting crust	old	⟷	young
(*d*) subduction of ridge	yes		—
(*e*) collision of arc (micro)continent, etc.	(yes)		yes
(*f*) crust on upper plate	thickness inhibits		—
(*g*) arc polarity	east facing	⟷	west facing

controls

(i) vector of relative plate motion (*a*, *b*)
(ii) nature and configuration of lower plate (*c*, *d*, *e*)
(iii) nature and configuration of upper plate (*f*)
(iv) absolute plate motion; ? Earth rotation (*g*)
(v) time

(1) The factors controlled by the vector of relative plate motion, convergence rate and convergence direction can clearly influence the state of stress in the upper plate, both to enhance back-arc basin opening and to enhance basin closure and destruction. A low convergence rate and oblique subduction both would tend to lower the compressive stress on the upper plate and hence encourage back-arc extension and basin formation. A high convergence rate and direct convergence both would tend to do the opposite. Hence changes in relative plate motion can lead to basin opening after a period of compression in the upper plate, or to basin closure and destruction after a basin formed. A situation may also arise where the relative plate motion would encourage maintenance of a particular situation, whether it be the existence of a previously formed basin or the existence of a basin-free central Andean type of margin.

For southern Chile the most obvious possibility for a change in relative plate motion to have influenced the situation lies in the timing of basin closure. The change from extensional (basin forming) deformation to compressional (basin destroying) deformation can be bracketed as between the late Early Cretaceous and the middle Late Cretaceous, and possibly refined to the Cenomanian. This is very close to the time at which the relative plate motion between the South American and African plates changed (Rabinowitz & LaBreque 1979) and a worldwide increase in spreading rates occurred (Larson & Pitman 1972).

Less concrete is the possibility that the Late Jurassic opening of the back-arc basin may have been enhanced by oblique subduction near the southern limit of the South American plate as Gondwanaland started to break up. Given the tectonics of the Andaman Sea, Bransfield Strait and the proto-Gulf of California, there is at least a suggestion that back-arc basins in west-facing arc systems need some unusual circumstances such as oblique subduction or the subduction of a 'leaky transform' or spreading ridge to initiate them (Curray *et al.* 1980).

(2) Turning to the configuration of the downgoing plate, Molnar & Atwater (1978) have pointed out the potential effect of subducting old dense lithosphere as opposed to young buoyant lithosphere. Uyeda & Kanamori (1979) and Dewey & Windley (1981) have also discussed this effect. Like changes in relative plate motion, it would appear that changes in the density of the downgoing slab could lead to back-arc basin opening along a margin previously under compression due to subduction of buoyant lithosphere, or vice versa if the change was towards younger (less dense) crust being subducted beneath an established basin. In the southern Andean example there is little that can be said here except to point out that before the intervention of the Chile Rise the crust being subducted beneath the southern Pacific margin of South America would presumably have been older and denser than that part of the Nazca plate currently being subducted beneath the central Andes and would therefore have favoured back-arc spreading.

It has been suggested that subduction of a spreading ridge can lead directly to back-arc basin formation (Uyeda & Miyashiro 1974; Uyeda & Kanamori 1979). There is no way of telling whether or not this happened along the southern South American margin in the Late Jurassic–Cretaceous. Again, however, this is the sort of mechanism that could overcome any general influence inhibiting back-arc basin formation in a west-facing subduction system. Unlike the other factors considered so far, it does not seem likely that the influence of this factor is reversible; nor is it independent, of course. Subducting a spreading ridge will inevitably lead to changes in relative plate motion and has been suggested as the cause of basin and range extension in North America (Atwater 1970; Scholz *et al.* 1971).

Collision of an arc, aseismic ridge, microcontinent or continent on the downgoing slab with the upper plate (Kelleher & McCann 1977) is clearly a way to increase the compression above the subduction zone and destroy a back-arc basin. Moreover, while not apparently a 'reversible' process, the collision of aseismic ridges may even result in localized oblique convergence and enhance the opening of back-arc basins. There is no evidence to suggest that this type of situation influenced the opening or closing of the back-arc basin in the southern Andes.

(3) It seems likely that the thickness and buoyancy of the crust forming the upper plate at convergent margin is a factor influencing the facility with which a back-arc basin opens (Toksöz & Bird 1977). All other factors being equal, a very thick and buoyant upper plate may tend to inhibit basin formation, although it obviously will not tend to destroy one that has already developed. The pre-Middle Jurassic basement of the southern Andes consists mainly of a fore-arc accretionary complex and the crust is only about 30 km thick (Ludwig *et al.* 1965). This may have made basin opening comparatively easy, certainly compared with the situation in the Central Andes.

(4) The question of absolute plate motion, whether or not related to Earth rotation, is a

complex one. Evidence around the southern rim of the Pacific Ocean basin seems to hold a tantalizing hint that extensional basins may have been more common along the western boundary of the ocean in the Palaeozoic and Early Mesozoic as well as in the Late Mesozoic and Cainozoic. This would suggest that throughout the Phanerozoic, back-arc extension has tended to be associated with east-facing arc systems. Obviously, if such an influence exists, the southern Andean basin opened in spite of it as a result of other factors operating at the time, such as oblique subduction or subduction of a spreading ridge.

(5) Finally, it seems intuitively obvious that time plays an important role in back-arc basin formation. The longer a factor tending to enhance back-arc basin formation stays in operation, the better will be the chances that such a basin actually forms, overcoming such obstacles as thick, buoyant crust on the upper plate (Toksöz & Bird 1977). Yet all small marginal ocean basins above subduction zones seem to be comparatively short-lived. It seems as though either the process of arc evolution tends to result in a shift of the locus of extensional tectonics with time, as in the western Pacific, or else the balance in favour of back-arc basin opening is so slight that it can readily be upset by one of the many factors that can lead to compression in the upper plate and hence to basin destruction. It does appear that we must conclude that the odds do somehow seem to be loaded against back-arc basin formation in west-facing arc systems.

Field work in southern South America and on South Georgia by geologists based at Lamont–Doherty Geological Observatory has been supported by the U.S. National Science Foundation mainly through the Division of Polar Programs, but also through the Earth Science Division, the International Decade of Oceanographic Exploration and the Office of International Programs. Preparation of this paper was funded through the following grants: DPP 78-20629, EAR 79-06308 and INT 79-20213.

I am particularly grateful to Richardson Allen, Don Elthon and Ian Ridley of Lamont–Doherty, and to Geoff Tanner, Brian Storey, David Macdonald and the British Antarctic Survey, for permission to reproduce their maps of the Sarmiento and Tortuga complexes and of South Georgia before publication, as well as for their helpful comments. Louis Agguire, Wallace Pitcher and John Tarney offered constructive criticism. Discussion with Joseph Curray on the origin of back-arc basins was valuable.

The figures were drafted by Linda Clark Dalziel.

REFERENCES (Dalziel)

Allen, R. B. 1980 *Antarct. Jl U.S.* (In the press.)

Andrews-Speed, C. P. 1980 *Geol. Mag.* **117**, 339–349.

Anon. 1980 *British Antarctic Survey Annual Report, 1978–79.* (75 pages.) Cambridge.

Atwater, T. 1970 *Bull. geol. Soc. Am.* **81**, 3715–3722.

Barker, P. F., Dalziel, I. W. D., Dinkelman, M. G., Elliott, D. H., Gombos, Jr, A. M., Lonardi, A., Plafker, G., Tarney, J., Thomson, R. W., Tjalsma, L. C., von der Borch, C. C. & Wise, Jr, S. W. 1976 In *Initial reports of the Deep Sea Drilling Project*, **36**, 993–1014. Washington, D.C.: U.S. Government Printing Office.

Bell, C. M., Mair, B. F. & Storey, B. C. 1977 *Br. Antarct. Surv. Bull.* no. 46, pp. 109–127.

Bruhn, R. L. 1979 *Bull. geol. Soc. Am.* **90**, 998–1012.

Bruhn, R. L. & Dalziel, I. W. D. 1977 In *Island arcs, marginal basins and deep sea trenches* (*Maurice Ewing Series*, vol. 1) (ed. M. Talwani & W. C. Pitman III), pp. 395–405. Washington, D.C.: American Geophysical Union.

Bruhn, R. L., Stern, C. R. & de Wit, M. J. 1978 *Earth planet. Sci. Lett.* **41**, 32–46.

Chase, C. G. 1978 *J. geophys. Res.* **83**, 5385–5387.

Cox, K. G. 1978 *Nature, Lond.* **274**, 47–49.

Curray, J. R., Emmel, F. J., Moore, D. G. & Raitt, R. W. 1980 In *The ocean basins and margins* (ed. A. E. M. Nairn & F. G. Stehli), vol. 6 (*The Indian Ocean*). New York and London: Plenum.

Dalziel, I. W. D. 1974 In *The geology of continental margins* (ed. C. A. Burk & C. L. Drake), pp. 567–579. New York: Springer-Verlag.

Dalziel, I. W. D. 1981 In *Antaractic geoscience* (ed. C. Craddock). Madison: University of Wisconsin Press.

Dalziel, I. W. D. & Cortès, R. 1972 In *Proc. 24th Session, Int. Geol. Congress, Montreal*, § 3, pp. 316–327.

Dalziel, I. W. D., de Wit, M. J. & Palmer, K. F. 1974 *Nature, Lond.* **250**, 291–294.

Dalziel, I. W. D., Dott, Jr, R. H., Winn, Jr, R. D. & Bruhn, R. L. 1975 *Bull. geol. Soc. Am.* **86**, 1034–1040.

Dalziel, I. W. D. & Forsythe, R. D. 1977 *Geol. Soc. Am. Abstracts with Programs, 1978 Joint Annual Meeting*, p. 385.

Dalziel, I. W. D. & Palmer, K. F. 1979 *Bull. geol. Soc. Am.* **90**, 259–280.

Davis, G. H. & Coney, P. J. 1979 *Geology* **7**, 120–124.

Dewey, J. F. & Windley, B. F. 1981 *Phil. Trans. R. Soc. Lond.* A **301**. (In the press.)

de Wit, M. J. 1977 *Tectonophysics* **37**, 53–81.

de Wit, M. J. & Stern, C. 1976 *Nature, Lond.* **264**, 615–619.

de Wit, M. J. & Stern, C. 1978 *J. Volcanol. geotherm. Res.* **4**, 55–80.

Dott, Jr, R. H., Winn, Jr, R. D., de Wit, M. J. & Bruhn, R. L. 1977 *Nature, Lond.* **266**, 620–622.

Elthon, D. 1979 *Nature, Lond.* **278**, 514–518.

Elthon, D. 1980 In *The sea* (ed. C. Emiliani), vol. 7. (In the press.)

Elthon, D. & Ridley, W. I. 1980 In *Proc. Int. Ophiolite Symposium*, Cyprus 1979. (In the press.)

Elthon, D. & Stern, C. 1978 *Geology* **6**, 464–468.

Forsyth, D. W. 1975 *J. geophys. Res.* **80**, 1429–1443.

Forsythe, R. D. & Mpodozis, A. C. 1979 *Revta geol. Chile*, no. 7, p. 13–29.

Gansser, A. 1973 *J. geol. Soc. Lond.* **129**, 93–131.

Herron, E. M. & Tucholke, B. E. 1976 In *Initial reports of the Deep Sea Drilling Project* (ed. C. D. Hollister, C. Craddock *et al.*), pp. 263–273. Washington, D.C.: U.S. Government Printing Office.

Hervè, F. *et al.* 1979 *Revta geol. Chile* **7**, 31–40.

Jurdy, D. M. 1979 *J. geophys. Res.* **84**, 6796–6802.

Karig, D. E. 1971 *J. geophys. Res.* **76**, 2542–2561.

Katz, H. R. 1973 *Jl R. Soc. N.Z.* **3**, 333–362.

Katz, H. R. & Watters, W. A. 1966 *N.Z. Jl Geol. Geophys.* **9**, 323–359.

Kelleher, J. & McCann, W. 1977 In *Island arcs, marginal basins and deep sea trenches* (*Maurice Ewing Series*, vol. 1) (ed. M. Talwani & W. C. Pitman III), pp. 115–122. Washington, D.C.: American Geophysical Union.

Larson, R. L. & Pitman III, W. C. 1972 *Bull. geol. Soc. Am.* **83**, 3645–3661.

Ludwig, W., Ewing, J. & Ewing, M. 1965 *J. geophys. Res.* **70**, 1855–1876.

Molnar, P. & Atwater, T. 1978 *Earth planet. Sci. Lett.* **41**, 330–340.

Natland, M. L., Gonzalez, E., Canon, A. & Ernst, M. 1974 *Mem. geol. Soc. Am.* no. 139. (126 pages.)

Nelson, E. P., Dalziel, I. W. D. & Milnes, A. G. 1980 *Eclog. geol. Helv.* **73**, 727–751.

Rabinowitz, R. D. & LaBrecque, J. 1979 *J. geophys. Res.* **84**, 5973–6002.

Saunders, A. D., Tarney, J., Stern, C. R. & Dalziel, I. W. D. 1979 *Bull. geol. Soc. Am.* **90**, 237–258.

Scholz, C. H., Barazangi, M. & Sbar, M. L. 1971 *Bull. geol. Soc. Am.* **82**, 2979–2990.

Scott, K. M. 1966 *Bull. Am. Ass. Petrol. Geol.* **50**, 72–107.

Simpson, P. & Griffiths, D. H. 1981 In *Antarctic geoscience* (ed. C. Craddock). Madison: University of Wisconsin Press. (In the press.)

Stern, C. R., de Wit, M. J. & Lawrence, J. R. 1976 *J. geophys. Res.* **81**, 4370–4380.

Stern, C. & Elthon, D. 1979 *Tectonophysics* **55**, 179–213.

Storey, B. C., Mair, B. F. & Bell, C. M. 1977 *Geol. Mag.* **114**, 203–208.

Suárez, M. 1976 *Geology* **4**, 211–214.

Suárez, M. 1979 *Geol. Mag.* **116**, 181–190.

Suárez, M. & Pettigrew, T. H. 1976 *Geol. Mag.* **113**, 305–328.

Tanner, P. W. G. 1981 In *Antarctic geoscience* (ed. C. Craddock). Madison: University of Wisconsin Press. (In the press.)

Tanner, P. W. G. & Rex, D. C. 1979 *Geol. Mag.* **116**, 167–179.

Tanner, P. W. G., Storey, B. C. & Macdonald, D. I. M. 1980 *Br. Antarct. Surv. Bull.* (In the press.)

Tarney, J., Dalziel, I. W. D. & de Wit, M. J. 1976 In *The early history of the Earth* (ed. B. F. Windley), pp. 131–146. London: Wiley.

Toksöz, M. N. & Bird, P. 1977 In *Island arcs, marginal basins and deep sea trenches* (*Maurice Ewing Series*, vol. 1) (ed. M. Talwani & W. C. Pitman III), pp. 379–394. Washington, D.C.: American Geophysical Union.

Uyeda, S. & Kanamori, H. 1979 *J. geophys. Res.* **84**, 1049–1061.

Uyeda, S. & Miyashiro, A. 1974 *Bull. geol. Soc. Am.* **85**, 1159–1170.

Wilson, J. T. & Burke, K. 1972 *Nature, Lond.* **239**, 448–449.

Winn, Jr., R. D. 1978 *Bull. geol. Soc. Am.* **89**, 533–547.

Winslow, M. A. 1980 In *Thrust and nappe tectonics*. Geological Society of London. (In the press.)

Winslow, M. A. 1981 In *Antarctic geoscience* (ed. C. Craddock). Madison: University of Wisconsin Press. (In the press.)

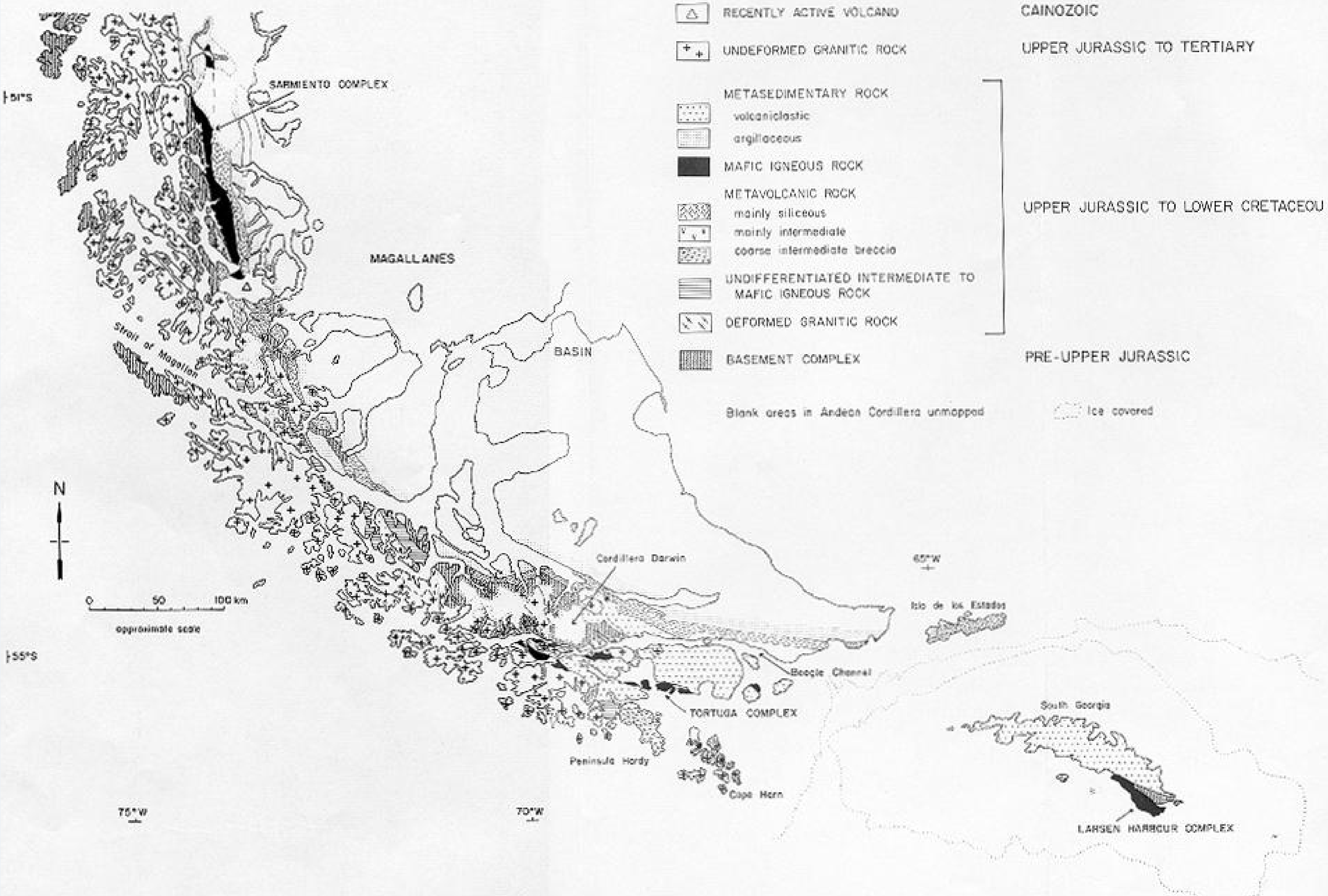


Figure 2. The geology of the southernmost Andes and South Georgia. The South Georgia platform is shown in its assumed approximate pre-drift position with respect to South America (see Dalziel *et al.* 1975; Tanner 1981).