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Deep generation of magmatic gas on the Moon and implications for pyroclastic eruptions

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[1] Lunar pyroclastic beads are interpreted to represent primitive magmas derived from great depths and rapidly erupted to the surface in explosive events. However, a detailed mechanism for gas generation at great depth and rapid magma transport to the surface has not yet been described. Furthermore, the pyroclastic beads are not petrogenetically related to basalts erupted near the sampling sites. We propose a model in which these conundrums are resolved through gas build-up in a low-pressure micro-environment near the tip of a magma-filled crack (dike) propagating rapidly from the magma source depth to the surface. The gas rich region consists of a free gas cavity overlying a foam extending vertically for ~20 km. Eruption of the foam results in the widespread emplacement of unfractionated pyroclastic beads. Subsequent ascent of the underlying gas-free picritic magma is unlikely to occur, perhaps accounting for the lack of sampled eruptive equivalents. *INDEX TERMS:* 5480 Planetology: Volcanism (8450); 6250 Planetology: Solar System Objects: Moon (1221); 8414 Volcanology: Eruption mechanisms. **Citation:** Wilson, L., and J. W. Head III, Deep generation of magmatic gas on the Moon and implications for pyroclastic eruptions, *Geophys. Res. Lett.*, *30*(12), 1605, doi:10.1029/2002GL016082, 2003.

1. Introduction

[2] Lunar dark mantling deposits are interpreted to represent pyroclastics occurring in a range of volcanic eruptive environments. Apollo astronauts collected submillimeter volcanic beads that are interpreted to be produced by explosive volcanic eruptions [e.g., *Heiken et al.*, 1974; *Sato*, 1979]. The occurrence of volcanic glasses in all Apollo soils [*Delano*, 1986] illustrated the widespread nature and significance of these materials. Although the Moon is deficient in volatiles, CO-rich gas produced by graphite oxidation [*Fogel and Rutherford*, 1995] is considered the most likely volatile driving the explosive eruptions that emplaced these types of deposits [*Wilson and Head*, 1981].

[3] Early work on the nature and pressure-dependence of reactions involving graphite [*Sato*, 1979] led to the interpretation that CO release occurred only in the upper several kilometers of the lunar crust during the final stages of magma ascent in dikes from subcrustal depths [*Wilson and Head*, 1981]. Analysis of the volcanic beads showed, however, that they largely represented primitive picritic compositions probably derived from great depths (>360 km) and retaining their primary compositions during ascent from their source regions to the surface, over several hundreds of kilometers distance [see review in *Longhi*, 1992]. This raised several fundamental questions: 1) how could sufficient gas be generated at such great depths and pressures to cause magmatic foams that would ultimately lead to disruption at the surface and pyroclast bead formation, and 2) how could magmas produced at depths in excess of 360 km be transported to the surface rapidly enough to maintain their primary compositions?

[4] Previous workers suggested several possible explanations generally rejecting them. *Longhi* [1992] and *Spera* [1992] both pointed out the need for very rapid magma transport. *Longhi* [1992] showed that given the volatile-depleted nature of the Moon, saturating a magma at these depths with a volatile species is very difficult.

[5] In this analysis, we propose a model in which this conundrum is resolved through gas build-up in the low-pressure micro-environment at the dike tip which is rapidly propagating from the magma source region to the surface. Gas generation occurs at great depths and throughout magma ascent; magma is transported at great speed, sufficient to avoid fractionation; and penetration of the dike to the surface results in widespread areal distribution of pyroclastic beads.

2. Gas Build-Up in the Tips of Rising Dikes

[6] The pressure gradient near the tip of a propagating dike must always be very large in order to force magma into the narrow tip, and so the pressure in the tip can be extremely small – in theory zero – even when the tip is at a great depth below the planetary surface. In practice, for eruptions on Earth, the exsolution of magmatic volatiles commonly buffers the tip pressure at some finite value [*Lister*, 1990a; *Lister and Kerr*, 1991; *Rubin*, 1993], and the tip space is occupied by a pocket of the most soluble volatile species. Immediately behind this is a foam of bubbles of gas, both the most soluble gas and other, less soluble species in the melt near the dike tip (Figure 1). In the lunar case it is generally accepted that the major gas species present was the CO produced by reactions between free carbon in the melt and oxides of Fe, Cr and Ti [*Fogel and Rutherford*, 1995]: The lunar magmas giving rise to the green and yellow pyroclastic glasses had the potential to produce a mass fraction M of at least ~1000 ppm (i.e. 10^{-3}) of CO. There is evidence that other volatiles, chiefly S, would have been partitioned into the gas phase, possibly adding at least another 500 ppm by mass [*Hess*, 2003, citing data in *Delano et al.*, 1994 and *Weitz et al.*, 1997].

[7] The CO production reaction proceeds at pressures less than a critical value determined by the oxygen fugacity

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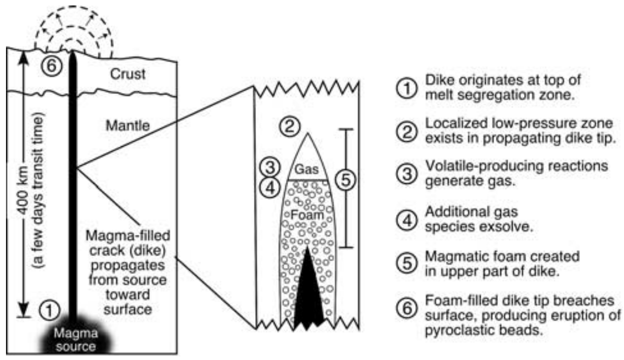

Figure 1. Processes operating in the low-pressure micro-environment of a dike tip, and the generation, rapid ascent and eruption of magmatic foams.

of the graphite oxidation surface [*Fogel and Rutherford*, 1995]. Estimates of the oxygen fugacity in the lunar mantle vary widely, but *Fogel and Rutherford* [1995] suggest that the critical pressure is likely to be 90 to 110 MPa (900 to 1100 bars). Thus the pressure in a lunar dike tip should be close to or less than this value, the exact pressure depending on the ability of the magma to deliver gas to the tip and the size of the tip, both determined by the dynamics of dike propagation. Where equilibrium cannot be maintained, and the tip pressure is less than the critical value, gas bubbles will form throughout a finite vertical region in the magma below the tip [*Lister*, 1990a].

3. Models of Lunar Dike Propagation

[8] Several treatments exist for the initial propagation of a dike away from a magma source region. *Lister* [1990b] and *Rubin* [1993] treated a dike containing neutrally buoyant magma under excess pressure. *Rubin*'s [1993] treatment is 2-dimensional and applies to both horizontal and vertical magma flow, whereas *Lister*'s [1990b] treatment is 3-dimensional and specific to lateral flow along a neutral buoyancy level. *Lister* [1990a] modelled the vertical rise of buoyant magma from both a point source and a line source, unpressurized in each case; and *Lister and Kerr* [1991] reviewed and extended aspects of the treatments. There is no published analytical 3-dimensional treatment of the vertical rise of magma (whether buoyant or not) from a pressurized source - an unfortunate situation given that this condition is encountered in all eruptions on Earth.

[9] In the case of the Moon, additional issues arise. Petrologic evidence indicates that all lunar magmas originated deep within the mantle, with those linked to the majority of pyroclastic glass beads coming from the greatest depths. These magmas will have been positively buoyant in their source zones [though possibly only marginally so - *Hess*, 2003], and it is not likely that the source zones will have been heavily pressurized since the deep mantle should not be able to support significant deviatoric stresses. Thus if the magmas remained positively buoyant throughout their rise to the surface, the models of *Lister* [1990a] should suffice to describe their ascent. However, the low density of the lunar crust implies that magmas should be negatively buoyant over much of their ascent through the crust. Exactly where the level of neutral buoyancy lies for a given magma composition and hence density is uncertain, though improved gravity data from recent lunar missions imply greater crustal densities [*Wieczorek et al.*, 2001] than were inferred in the past [e.g. *Wilson and Head*, 1981]. We approach these issues as follows.

[10] The model of *Rubin* [1993] treats dikes that are driven by an excess pressure in the magma source zone. We assume that picritic magmas originate at a depth D = 400 km in the lunar mantle, this value lying within the 360 to >500 km range inferred by *Longhi* [1992]. The melt is assumed to be locally buoyant with a density difference relative to the host rocks at its source of $\Delta\rho_s = 150$ kg m^{-3} (*Hess* [2003] argues that the values assumed in the past, $\sim$300 kg m^{-3}, may be too large). If the source zone, presumably the upper part of a rising convection cell, has a vertical extent Z of, say, $\sim$20 km then the magma source will have an effective driving pressure P_d (= g Z $\Delta\rho_s$ where g is the acceleration due to gravity, 1.63 m s^{-2}) of about 5 MPa, and we use this value in what follows. The low-viscosity magma is treated as rising in a dike rather than as a diapir on the basis of arguments about the effects of the viscosity contrast with the host rocks [*Rubin*, 1993].

[11] The dike propagation analysis of *Rubin* [1993] relates the length, R, of the gas-filled cavity in the dike tip to the dike length L, the tip pressure P_{tip}, the dike driving pressure P_d, the host rock stiffness $G = [\mu/(1 - \nu)]$ where μ is the shear modulus and ν is the Poisson's ratio, the dimensionless ratio $[(3\ \eta_m)/(2\ \eta_r)]^{1/3}$ where η_m is the viscosity of the magma ($\sim$1 Pa s -) and η_r is the viscosity of the host rocks, and the local least principle stress (approximately the lithostatic load) in the host rocks P_{lith}. We adopt G = 10 GPa as typical of rocks forming planetary lithospheres [*Rubin*, 1993] and note that for all reasonable estimates of lunar magma viscosity and mantle viscosity the quantity $[(3\ \eta_m)/(2\ \eta_r)]^{1/3}$ is less than 10^{-6} and has a negligible influence on what follows. We fix P_{tip} at 100 MPa on the basis of the chemical arguments about the CO production reaction discussed above and can then use Figure 7a in *Rubin* [1993] to find the ratio R/L as a function of the quantity $S = [(P_{tip} - P_{lith})/P_d]$. At great depth P_{lith} is very much greater than P_{tip} and S is very negative. As the dike tip ascends, P_{lith} decreases, and S becomes less negative. At some finite depth (just deeper than 21 km for P_{tip} = 100 MPa) S becomes zero, and as the dike tip approaches this depth the gas cavity becomes very large. As discussed at length by *Lister* [1990a] and *Lister and Kerr* [1991], the dynamics of gas accumulation in dike tips is so complex that analytical treatments of the process are very likely to be unreliable. Nevertheless we can use *Rubin*'s [1993] treatment to estimate the depth at which conditions become critical. (Figure 2a) shows the length R_{crit} of the gas-dominated region at the point where the dike tip is about to break through to the surface, calculated using the rock and magma properties specified above. R_{crit} depends only weakly on the value assumed for the dike driving pressure P_d and is about 22 km for the most likely gas-filled tip pressure, P_{tip} = 100 MPa, being essentially proportional to the value assumed.

[12] In the model of *Lister and Kerr* [1991] it is the buoyancy of the magma that drives its rise to the surface and it is not clear that lunar magmas, especially the picrites giving rise to the pyroclastic beads, fulfill the requirement of

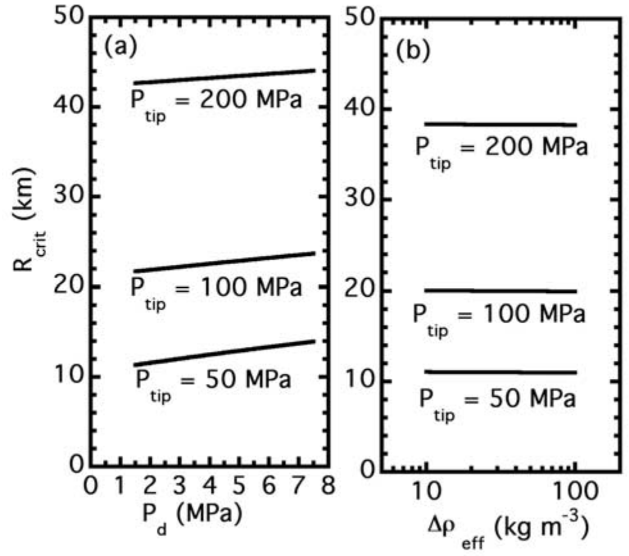

Figure 2. Vertical lengths of the gas-dominated upper parts of picritic dikes as a function of (a) driving pressure in the magma source zone using the model of *Rubin* [1993] and (b) magma buoyancy using the model of *Lister and Kerr* [1991]. The most likely gas pressure at the dike tip is 100 MPa but two other values are shown for comparison.

being buoyant at all levels in the crust. Given that we are dealing with the initial propagation of a dike assisted by gas exsolution we assume a net positive buoyancy corresponding to a net density difference $\Delta\rho_{eff}$; however, to allow for the possibility that this difference is small, we carried out the calculations for $\Delta\rho_{eff}$ = 10, 30 and 100 kg m^{-3}. *Lister and Kerr*'s [1991] models give the variation of the length and width of the dike in the horizontal plane as a function of height above the magma source. We find that for all conditions appropriate to dikes on the Moon the low viscosity of the magma causes the motion to be turbulent so that *Lister and Kerr*'s equations (58a) and (58b) apply; we then use their equation (55) to find the approximate length of the gas-dominated dike tip region [our R_{crit} is their s_{vol}, rendered dimensional using the scale length given by their equation (31a)]. To use these equations it is necessary to choose a value for the total volume flux, Q, of magma flowing through the dike. We examined the range of mare lava eruption rates deduced from the lengths of lava flows, the lengths of sinuous rilles, and the source geometries of sinuous rilles: values of Q lie in the range 20,000 to 200,000 m^3 s^{-1}. At the lower end of this range, a check on the thermal viability of the resulting dikes using equations (21–23) of *Wilson and Head* [1981] shows that mare lava magma sources would have to be at depths less than 200 km to avoid excessive magma cooling during transport through the dike. The sources of the picritic glasses must be at greater depths than this, and a depth of 400 km requires a minimum flux of about 70,000 m^3 s^{-1}. We used Q = 100,000 m^3 s^{-1} in the bulk of the calculations. Figure 2b shows the variation of R_{crit} with $\Delta\rho_{eff}$ for the same three values of P_{tip} used in Figure 2a. Changing Q by a factor of 2 produces only a ~1% change in the values of R_{crit}.

[13] Comparison of Figures 2a and 2b shows that the values of R_{crit} given by *Lister and Kerr*'s [1991] model are remarkably similar to those from *Rubin*'s [1993] model (actually ~10% smaller), despite the differences in assumptions about the conditions driving the eruptions. Our estimate of P_{tip} = 100 MPa would imply R_{crit} = ~20 km.

[14] As pointed out by *Rubin* [pers. comm., 2002] all existing treatments of gas concentration in dike tips neglect subtleties related to the stability of the gas-rich region which may lead to an increase in R_{crit} and a decrease in P_{tip} (assumed constant in our analysis) as the dike tip nears the surface. However, this effect would have minimal implications for our analysis unless it led to an order of magnitude change in R_{crit}, which seems unlikely.

4. Discussion

[15] As a dike propagates upward, the magma flow field near the base of the region already dominated by gas is complex; as the dike tip passes beyond a given level the dike widens to a maximum value and magma very near the base of the gas-dominated region travels horizontally rather than vertically to accomodate the lateral widening (Figure 1). The dynamics of the nucleation of bubbles of exsolving gas in this region, and of the rise of the bubbles through the liquid close to the interface, has not been modeled in detail by previous authors and is not addressed here. However, the speed at which small (~10–20 μm diameter at nucleation) gas bubbles can rise through even a low-viscosity magma (~0.1 μm s^{-1}) and the speed at which magma might drain from between such bubbles as they become tightly packed (~5 μm s^{-1}), are both much less than the rise speed of the magma as a whole. Magma rise speeds are ~0.5 m s^{-1} for lunar picrites in the *Lister and Kerr* [1991] model, implying that ascent from the mantle source requires several days, and ~5 m s^{-1} in the *Rubin* [1993] model, implying a rise time of ~1 day. Thus as it approaches the surface, the gas-dominated tip region is likely to contain a very densely-packed foam rather than free gas.

[16] We now estimate the mass of gas per unit length of the dike along strike which is contained in the near-tip region. In *Rubin*'s [1993] model (his Figure 6a) we can approximate the shape of the gas-dominated region as being similar to that of the dike. Thus if W_{crit} is the width of the base of the region of length R_{crit} we have $(W_{crit}/R_{crit}) = (W_0/L)$ where W_0 is the width of the dike at the magma source level, and (W_0/L) is given by *Rubin*'s Figure 4 as being slightly greater than P_d/G. Using G = 10 GPa and P_d = 2.5 MPa, (W_0/L) = ~2.5 × 10^{-4}. With L = 400 km, W_0 = ~100 m, and since R_{crit} is ~20 km, W_{crit} would be ~5 m. The approximate area of the tip region is then that of a triangle of base 5 m and height 20 km, i.e. 5 × 10^4 m^2. On the same basis the total cross-sectional area of the dike is 2 × 10^7 m^2. Thus the gas-dominated region occupies 0.25% of the dike area. The density of the dike magma is ~3000 kg m^{-3} and the density of the gas is found from the gas law using the pressure, 100 MPa, the temperature (we assume that the gas is in equilibrium with the magma at ~1750 K) and the mean molecular weight of a mixture of CO and sulphur gases, say ~35 kg kmol^{-1} as ~240 kg m^{-3}. Thus, if the gas-dominated region is a closely packed foam with gas occupying ~90% of the volume, the gas represents ~200 ppm of the magma mass in the dike. The total potential gas content of all the magma in the dike is, as we saw earlier, ~1500 ppm, and

so only ∼13% of the dike magma needs to be able to segregate gas into the tip region to explain its size. It is not easy to estimate the width of the gas-dominated region in the case of the model in *Lister and Kerr* [1991], though *Lister*'s [1990a] Figure 6 implies that it is typically a factor of 2 or 3 narrower than would be predicted by *Rubin*'s [1993] model, requiring even less of the dike magma to transfer gas to the region.

[17] When the tip of a dike breaks through the lunar surface, free gas at the tip escapes very quickly, followed by the consequences of exposing the ∼20 km long underlying foam region, containing gas bubbles at a pressure of ∼100 MPa, to a vacuum. An expansion wave travels downward into the foam at about half the speed of sound in the foam, i.e. ∼50 m s^{-1} [*Kieffer*, 1977; *Wilson and Head*, 1981], disaggregating the foam into free gas entraining melt droplets which are accelerated upward by the gas expansion into the overlying vacuum. Irrespective of the exact gas pressure, under these conditions the droplets leave the vent region with almost all of the ultimate speed reached by the gas, $U = \{(n\ Q\ T\ \gamma)/[m\ (\gamma - 1)]\}^{1/2}$, where n is the gas mass fraction of gas in the foam (∼0.4 if the foam is 90% by volume gas with the density ∼240 kg m^{-3} found above), γ = ∼1.4 and m = ∼35 kg $kmol^{-1}$ are the ratio of the specific heats and the molecular weight, respectively, of the gas, Q = 8.314 kJ $kmol^{-1}$ is the universal gas constant and T = ∼1750 K is the gas temperature. Thus U is ∼760 m s^{-1} and the maximum distance to which glass beads might be ejected, if they leave the vent region with an elevation of 45°, is ∼350 km. It seems likely that as the deeper parts of the gas-rich region of the dike are emptied the spray emerging from the vent will become more collimated, thus reducing the range, though disruption of the surrounding regolith and abrasion of the vent walls may partly compensate for this. Division of the volume of the foam region by the area of dispersion shows that depending on the depth to which foam is excavated, the thicknesses of layers of pyroclastic beads might be expected to range from less than a few cm to ∼30 cm.

[18] It seems unlikely that degassed magma below the explosively excavated region would reach the surface. The expansion wave travels into the foam at ∼50 m s^{-1}, an order of magnitude faster than the magma is rising through the dike, and leads to a great increase in the stress acting across the wall of the dike, ultimately reaching ∼95 MPa at the maximum ∼20 km depth. Thus failure of the dike walls and choking of the dike is likely to occur, possibly before the gas-dominated region has been completely emptied, and since the degassed magma is almost certainly substantially denser than the shallow crust [*Wieczorek et al.*, 2001] there is no reason to expect any significant effusive eruption: the bulk of the picritic magma remains at depth as an intrusion.

5. Implications for Origin and Eruption of Primitive Picritic Beads

[19] We have shown that: 1) gas can be generated at great depths in the lunar interior (e.g., >360 km, the inferred depths of origin of primitive picritic magmas interpreted to be parent to the pyroclastic beads); 2) the gas-generation micro-environments are the low pressure zones at the tips of propagating dikes as they set out from magma source regions; 3) sufficient gas can be generated to produce gas-filled cavities and magmatic foams; 4) these will explosively decompress upon arrival at the surface to produce widespread pyroclastic bead deposits; 5) these deposits would commonly extend over areas up to ∼700 km wide; 6) the transport times of magma to the surface are very rapid (of order one to several days), sufficient to avoid significant fractionation; and 7) there is no reason to expect associated effusive eruptions of the picritic magmas giving rise to the pyroclastics. This process thus provides a plausible mechanism for: 1) the generation and rapid delivery to the surface of gas-rich magmas from great depths within the Moon; 2) the lack of correlation between the pyroclastic beads and the basalts sampled at the Apollo landing sites; and 3) the apparently separate petrogenesis of pyroclastic beads and effusive volcanic deposits.

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References

Delano, J., Pristine lunar glasses: Criteria, data, and implications, *J. Geophys. Res.*, *91*, D201–D213, 1986.

Delano, J. W., B. Z. Hanson, and E. B. Watson, Abundance and diffusivity of sulphur in lunar picritic magmas, *Proc. Lunar Planet. Sci. Conf*, 25th, 325–326, 1994.

Fogel, R., and M. Rutherford, Magmatic volatiles in primitive lunar glasses, I, FTIR and EPMA analyses of Apollo 15 green and yellow glasses and revision of the volatile-assisted fire-fountain theory, *Geochim. Cosmochim. Acta*, *59*, 201–215, 1995.

Heiken, G. H., D. S. McKay, and R. W. Brown, Lunar deposits of possible pyroclastic origin, *Geochim. Cosmochim. Acta*, *38*, 1703–1718, 1974.

Hess, P., Mare magmatism, in *New Views of the Moon*, edited by B. Joliff and M. Wieczorek, Mineralogical Society of America Special Publication, in preparation, 2003.

Kieffer, S. W., Sound speed in liquid-gas mixtures: Water-air and water-steam, *J. Geophys. Res.*, *82*, 2895–2990, 1977.

Lister, J. R., Buoyancy-driven fluid fracture: The effects of material toughness and of low-viscosity precursors, *J. Fluid. Mech.*, *210*, 263–280, 1990a.

Lister, J. R., Buoyancy-driven fluid fracture: Similarity solutions for the horizontal and vertical propagation of fluid-filled cracks, *J. Fluid. Mech.*, *217*, 213–239, 1990b.

Lister, J. R., and R. C. Kerr, Fluid-mechanical models of crack propagation and their application to magma transport in dykes, *J. Geophys. Res.*, *96*, 10,049–10,077, 1991.

Longhi, J., Experimental petrology and petrogenesis of mare volcanics, *Geochim. Cosmochim. Acta*, *56*, 2235–2252, 1992.

Rubin, A. M., Dikes vs. diapirs in viscoelastic rock, *Earth Planet. Sci. Lett.*, *119*, 641–659, 1993.

Sato, M., The driving mechanism of lunar pyroclastic eruptions inferred from the oxygen fugacity behavior of Apollo 17 orange glass, *Proc. Lunar Planet. Sci. Conf.*, *10th*, 311–325, 1979.

Spera, F., Lunar magma transport phenomena, *Geochim. Cosmochim. Acta*, *56*, 2253–2265, 1992.

Weitz, C. M., M. J. Rutherford, and J. W. Head, Oxidation states and ascent history of the Apollo 17 volcanic beads as inferred from metal-glass equilibria, *Geochim. Cosmochim. Acta*, *61*, 2765–2775, 1997.

Wieczorek, M. A., M. T. Zuber, and R. J. Phillips, The role of magma buoyancy on the eruption of lunar basalts, *Earth Plan. Sci. Lett.*, *185*, 71–83, 2001.

Wilson, L., and J. W. Head, Ascent and eruption of basaltic magma on the Earth and Moon, *J. Geophys. Res.*, *86*, 2971–3001, 1981.

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