3.11 TURBULENT TRANSFER COEFFICIENTS AND ROUGHNESS LENGTHS OVER SEA ICE: THE SHEBA RESULTS

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1. INTRODUCTION

On the night of 30 January 1998, the ice cracked around the Canadian icebreaker Des Groselliers and severed the power lines than ran the SHEBA ice camp. Lights went out in the camp, and data collecting stopped at our main 20-meter Atmospheric Surface Flux Group (ASFG) tower for almost eight days. Despite this breakup and many others, the experiment to study the Surface Heat Budget of the Arctic Ocean (SHEBA) still produced the largest and most comprehensive set of surface-level meteorological and turbulence data ever collected over sea ice.

Andreas et al. (1999) and Persson et al. (2002) describe the instruments that our group deployed and maintained in and around the SHEBA camp. For example, besides that main tower, we always had three or four remote portable automated mesonet (PAM) stations deployed in the vicinity of the camp. These were Flux-PAM stations designed and built by NCAR’s Integrated Surface Flux Facility. Since these were self-contained, during the breakup, they continued recording despite the power outage in the main camp. Here we report analyses of the eddy-correlation measurements of the momentum and sensible heat fluxes that we made through the SHEBA year at our main tower and at these PAM stations.

The momentum flux or surface stress ($\tau$) and the sensible heat flux ($H_s$) couple the lowest levels of the atmosphere to the sea ice surface and, thus, serve as some of the boundary conditions for mesoscale and large-scale atmospheric models. Such models typically estimate these turbulent surface fluxes with a bulk flux algorithm formulated as

$$\tau = \rho C_{Dr} U_r^2,$$  \hspace{1cm} (1a)

$$H_s = \rho c_p C_{Hr} U_r (\Theta_s - \Theta_r).$$ \hspace{1cm} (1b)

Here, $\rho$ is the air density; $c_p$, the specific heat of air at constant pressure; $U_r$, the wind speed at reference height $r$; $\Theta_s$, the potential temperature at the surface; and $\Theta_r$, the potential temperature at height $r$.

The set (1) usually includes a third equation—one for the latent heat flux. But since the Flux-PAM stations did not measure the latent heat flux directly and since our eddy-correlation measurements of that flux on our main tower have, so far, proved inscrutable, we do not discuss parameterizations for the latent heat flux here. Instead, see Andreas (2002) or Andreas et al. (2003).

The secret to the bulk flux algorithm is evaluating the turbulent transfer coefficients in (1): $C_{Dr}$ is the drag coefficient appropriate at reference height $r$, and $C_{Hr}$ is the transfer coefficient at height $r$ for sensible heat. For example, NCAR’s Climate System Model (Bryan et al. 1996) uses a formulation like (1) over both open ocean and sea ice surfaces. $C_{Dr}$ and $C_{Hr}$ change with the surface, however. Here we use our SHEBA measurements to learn better how to estimate $C_{Dr}$ and $C_{Hr}$ over sea ice surfaces. We call our resulting synthesis the SHEBA bulk flux algorithm.
2. FORMAL BACKGROUND

As a consequence of Monin-Obukhov similarity theory, we can formally write $C_{Dr}$ and $C_{Hr}$ in (1) as

$$C_{Dr} = \frac{k^2}{\left[\ln\left(\frac{r}{z_0}\right) - \psi_m\left(\frac{r}{L}\right)\right]^2},$$
\hspace{1cm} (2a)

$$C_{Hr} = \frac{k^2}{\left[\ln\left(\frac{r}{z_0}\right) - \psi_m\left(\frac{r}{L}\right)\right] \ln\left(\frac{r}{z_T}\right) - \psi_h\left(\frac{r}{L}\right)}.$$ \hspace{1cm} (2b)

Here, $k$ is the von Kármán constant, which we take as 0.40 for our analysis, but see Andreas et al. (2002). Also in (2), $\psi_m$ and $\psi_h$ are "known" corrections for near-surface atmospheric stratification effects that depend on the stability parameter $\zeta \equiv r/L$, where $L$ is the Obukhov length. For unstable stratification ($\zeta < 0$), we use Paulson's (1970) functions for $\psi_m$ and $\psi_h$; for stable stratification ($\zeta > 0$), we use Holtslag and De Bruin's (1988) functions.

At our main tower and at the PAM stations, we measured all the quantities in (1) except $C_{Dr}$ and $C_{Hr}$. We can therefore evaluate these from the measurements. But, at least according to (2), the roughness lengths for wind speed ($z_0$) and temperature ($z_T$) seem to be more fundamental variables. And, in fact, developing theoretical expressions for $z_0$ and $z_T$ is often easier than treating $C_{Dr}$ and $C_{Hr}$ theoretically.

Consequently, having evaluated $C_{Dr}$ and $C_{Hr}$ from the data, we can invert (2) to obtain values for $z_0$ and $z_T$. That is,

$$z_0 = r \exp \left\{ -\left[ k C_{Dr}^{-1/2} + \psi_m(\zeta) \right] \right\},$$ \hspace{1cm} (3a)

$$z_T = r \exp \left\{ -\left[ k C_{Dr}^{1/2} C_{Hr}^{-1} + \psi_h(\zeta) \right] \right\}.$$ \hspace{1cm} (3b)

Realize that we also know $L$, and thus $\zeta$, from our measurements; therefore, computing $z_0$ and $z_T$ from our data is straightforward.

Once we know $z_0$ and $z_T$, we can compute the transfer coefficients in more standard form. To make comparing data from various sites and for various surfaces meaningful, we usually report $C_D$ and $C_H$ at a standard reference height of 10 m. Likewise, for such comparisons, we prefer to remove stratification effects to compare surface properties more fairly. From (2), we see that for neutral stability ($\psi_m = \psi_h = 0$) and for a 10-m reference height,

$$C_{DN10} = \frac{k^2}{\left[\ln(10/z_0)\right]^2},$$ \hspace{1cm} (4a)

$$C_{HN10} = \frac{k^2}{\left[\ln(10/z_0)\right] \left[\ln(10/z_T)\right]}.$$ \hspace{1cm} (4b)

These are the so-called neutral-stability transfer coefficients appropriate for a reference height of 10 m. Clearly, knowing $z_0$ and $z_T$ is equivalent to knowing the neutral-stability transfer coefficients.

3. THE ROUGHNESS LENGTH $z_0$

3.1 Overview of the Year

Our instruments ran at SHEBA from October 1997 through September 1998. All the data that we report here are based on hourly averages for this period.

Figure 1 shows a time series for the SHEBA year of $C_{DN10}$ values measured at our main tower and at the Flux-PAM stations called Atlanta, Baltimore, and Florida. The PAM stations made eddy-correlation measurements of $\tau$ and $H_s$ at one level only. Our main tower measured at five levels; all the tower values that we report here, however, are based on the median values of the available flux measurements from the five levels.

Because of riming problems with the sonic anemometer/thermometers on the PAM stations that were not corrected until March or April 1998, the data returns from these early in the experiment were spotty. Figure 1 therefore shows monthly averages of $C_{DN10}$ values from the PAM stations until March 1998 (SHEBA day 425 is 1 March 1998). In Fig. 1, for all the tower data and for the PAM stations starting in March, the plotted points are averages for the first 10 days of the month, for the second 10 days, and for the remainder of the month.

We identify two aerodynamic regimes in Fig. 1. From the start of the experiment through 14 May 1998 (day 499), the sea ice was snow covered, and the snow was dry enough to drift.
Figure 1. Neutral-stability, 10-m values of the drag coefficient from four SHEBA sites. Depending on the data available, we averaged hourly \( C_{DN10} \) values either monthly or by thirds of a month. Error bars are two standard deviations of the mean.

We call these conditions winter. These conditions resumed and winter thus returned (aerodynamically speaking) on 15 September 1998 (day 623). The \( C_{DN10} \) values during this period are quite scattered, which we suspect is a wind effect. (The scanty data from the PAMs early in this period could also explain the scatter, though.)

When the snow got sticky, stopped drifting, and eventually melted, however, the four sites depicted in Fig. 1 agree much better as to the value of \( C_{DN10} \). Surprisingly, when the snow was gone from the ice and the surface became pockmarked with melt ponds and leads, the sea ice appeared more aerodynamically homogeneous than a snow-covered surface. Hence, we treat the non-drifting and snow-free period, 15 May through 14 September 1998 (days 500–622), as an aerodynamic regime we call summer. Henceforth, we partition our data and analyses into winter and summer.

### 3.2 \( z_0 \) in Winter

Figure 2 shows bin-averaged \( z_0 \) values for winter (October 1997 through 14 May 1998 and 15 September to October 1998) measured on our main tower and at the three PAM stations with the most continuous records. The bins are 2 cm s\(^{-1}\) wide in \( u_0 \) for the smaller \( u_0 \) values and 5 cm s\(^{-1}\) wide for the larger values, where we had fewer observations.

This figure summarizes over 6,200 hourly \( z_0 \) values measured at the four sites. Most markers represent the average of many hourly \( z_0 \) values—sometimes over 200. Only a couple of markers in the highest \( u_0 \) bins are based on one or two values.

The curve in the figure is

\[
\frac{z_0}{u_0} = \frac{0.135 \nu}{u_0} + 2.0 \times 10^{-4} \exp\left(\left(\frac{u_0 - 0.25}{0.15}\right)^2\right) + \frac{0.03 u_0^2}{g},
\]

where \( \nu \) is the kinematic viscosity of air and \( g \) is the acceleration of gravity. Equation (5) gives \( z_0 \) in meters when the other variables are in MKS units.

Equation (5) is similar to expressions that Jordan et al. (2001) and Andreas et al. (2003) use to model \( z_0 \) for snow-covered sea ice on Ice Station Weddell. It suggests three aerodynamic regimes: an aerodynamically smooth regime where \( z_0 \) goes as \( \nu/u_0 \), a drifting snow regime where energy arguments (Owen 1964) suggest \( z_0 \) scales with \( u_0^2/g \), and an intermediate regime between these extremes where the “permanent” roughness of the surface dictates the momentum transfer. In essence, (5) is similar to the two-term summation that Smith (1988) and Fairall et al. (1996) use to model \( z_0 \) over the open ocean. Our result, however, has a third term, the middle term on the right, that accounts for the fundamental roughness of the surface.
We chose the coefficients in the second and third terms in (5) specifically to fit the $z_0$ data in Fig. 2. These tuning values are similar to but not exactly the same as in the expressions that Jordan et al. (2001) and Andreas et al. (2003) report. We are still studying the reasons for these differences.

The data in Fig. 2 do not seem well represented at low $u$ by the aerodynamically smooth relation, the first term on the right of (5). But measuring both $u$ and $z_0$ in very light winds is difficult because, in the Arctic, such conditions are usually variable, and the turbulence is often intermittent. We thus rely on wind tunnel measurements for this end of our parameterization.

### 3.3 $C_{DN10}$ in Summer

Figure 3 has three panels. The top panel repeats the $C_{DN10}$ values from Fig. 1 from the late winter through the end of the experiment. This series highlights the fairly consistent behavior of the $C_{DN10}$ values among the four sites during the summer.

As summer progressed at SHEBA, leads opened in the vicinity of the camp, all the snow melted, and the sea ice surface became increasingly pocked with melt ponds. The water features, in particular, created edges that the wind could push against. That is, form drag, which is sustained by pressure forces on the near-vertical ice faces (e.g., Banke et al. 1980; Andreas et al. 1984; Andreas 1995; Birnbaum and Lüpkes 2002), becomes a significant mechanism for air-surface momentum transfer in the summer. Our observations that the $C_{DN10}$ data in Fig. 1 converge better in summer than in winter suggest that form drag may dominate the momentum exchange in summer.

Since open water is associated with the vertical edges that created this form drag, in the middle panel of Fig. 3, we show the lead fraction of total surface area in the vicinity of the SHEBA camp, the melt pond fraction, and the total open water fraction (the sum of leads and ponds). These data are courtesy of D. K. Perovich (2001, personal communication). The $C_{DN10}$ data seem to follow the trend in the total water fraction. $C_{DN10}$ is low in mid-May 1998 when the ice is still snow covered. But $C_{DN10}$ begins increasing as open water appears in late May (day 517 is 1 June 1998) and peaks near the maximum in open water fraction. $C_{DN10}$ then returns to typical winter values as the open water freezes and becomes snow covered.

The peak in summer $C_{DN10}$ does not coincide perfectly with the peak in total water fraction. It is quite close, however, to the peak in pond fraction; but $C_{DN10}$ does not fall off beyond the peak as rapidly as the pond fraction does. Perhaps $C_{DN10}$ reflects a weighted sum of lead and pond coverages rather than just the total water fraction.

The bottom panel in Fig. 3 suggests that the areally averaged albedo may also be a predictor for $C_{DN10}$. These albedo values also come from D. K. Perovich (2001, personal communication) and derive from his measurements of surface types and their individual albedo values. The lower two panels in Fig. 3 suggest that we may be able to estimate $C_{DN10}$ from satellite observations.

We therefore pursue the relationship between summer $C_{DN10}$ and ice concentration (or equivalently, water fraction) in Fig. 4. Here we have interpolated the $C_{DN10}$ values in Fig. 3 to the corresponding water fraction, $W$, and, subsequently, to
the ice concentration, \( C = 1 - W \). Remember, \( C \) is the areal coverage of bare ice. There is, of course, ice at the bottom of the melt ponds that is not included in this value.

The data in Fig. 4 tend to support our contention that \( C_{DN10} \) increases as the ice concentration decreases from 100% and more edges emerge. But for SHEBA, the ice concentration was always greater than 60% (Figs. 3 and 4). Therefore, to complete the picture of how \( C_{DN10} \) depends on ice concentration, we add two more data sets to complement the SHEBA data in Fig. 4. These are observations in the Antarctic marginal ice zone near the Greenwich Meridian that Andreas et al. (1984) made and a more extensive aircraft survey in the Arctic marginal ice zone near Svalbard that Birnbaum and Lüpkes (2002) report. These two sets are not necessarily summer observations; but since we are primarily interested in edge effects related to open water, they do not have to be.

We fitted the observations in Fig. 4 with a second-order polynomial,

\[
10^3 C_{DN10} = 1.20 + 3.11C - 2.81C^2 , \tag{6}
\]

where \( C \) is the fractional ice concentration. Equation (6) suggests that \( C_{DN10} \) peaks at \( 2.06 \times 10^{-3} \) when \( C = 0.55 \). Realize, though, that this fit depends somewhat on our choice of endpoints—namely, the values of \( C_{DN10} \) over the open ocean (\( C = 0 \)) and over compact sea ice (\( C = 1 \)), which we took as \( 1.20 \times 10^{-3} \) and \( 1.50 \times 10^{-3} \), respectively. In essence, Fig. 4 shows a unified picture of how \( C_{DN10} \) depends on ice concentration, whether the ice surface is the marginal ice zone or the central pack in summer.

4. THE ROUGHNESS LENGTH \( z_T \)

4.1 Background

The roughness length for temperature, \( z_T \), in (2b) is analogous to \( z_0 \); it is the fictitious height at which the temperature profile meets the surface temperature \( \Theta_s \). From (4), we see that knowing \( z_0 \) and \( z_T \) is equivalent to knowing \( C_{DN10} \) and \( C_{HN10} \). But devising theoretical models for \( z_T \) is easier than modeling \( C_{HN10} \) directly (e.g., Brutsaert 1975; Liu et al. 1979; Andreas 1987). Hence, we focus our discussion on the behavior of \( z_T \) or \( z_T/z_0 \) since \( z_T/z_0 \) appears in

\[
C_{HN10} = \frac{C_{DN10}}{1 - k^{-1}C_{DN10}^2 \ln(z_T/z_0)} , \tag{7}
\]

which derives from (4).

Equation (7) shows that, when \( z_T = z_0 \), the neutral-stability transfer coefficients for momentum and sensible heat are the same. This equality is essentially the Reynolds analogy (e.g., Schlichting 1968, p. 268 f.). The stability-dependent transfer coefficients in (2) need not be the same, though, because the stratification corrections, \( \psi_m \) and \( \psi_h \), may be different.

Andreas (1987) developed a model for \( z_T/z_0 \) that is still the only theoretical prediction for how \( z_T \) behaves over surfaces of snow and ice. That model predicts \( z_T \) from

\[
\ln(z_T/z_0) = b_a + b_1 (\ln R_x + b_2 (\ln R_x)^2 , \tag{8}
\]

where \( R_x = u_z/v \) is the roughness Reynolds number. Andreas (1987) and Andreas (2002) tabulate the polynomial coefficients in (8) for flow that is aerodynamically smooth, in transition, or aerodynamically rough.

Andreas (2002) reviews some limited tests of the validity of (8) for both \( z_T \) and \( z_0 \), the roughness length for humidity. Andreas et al. (2003) report on how results from a more extensive data set, collected over snow-covered Antarctic sea ice, compare with (8) for both \( z_T \) and \( z_0 \). Here we describe how the SHEBA \( z_T \) data compare with (8).
4.2 $z_T$ in Winter

Figure 5 shows $z_T$ values measured on our main tower during the winter and plotted as (8) suggests. The data are quite scattered, but such scatter is typical in this business. After all, both $z_T$ and $z_0$ are exponential functions of the measurements they derive from [i.e., see (3)]. Nevertheless, the $z_T/z_0$ values tend to collect around the theoretical prediction, (8), and decrease with increasing $R_a$, as the theory predicts.

Andreas (2002), however, points out a problem with plots such as Fig. 5: Because $z_0$ appears on both axes, the plot is prone to fictitious correlation. To minimize this effect and to reduce the scatter, we replotted the $z_T$ values from Fig. 5 in Fig. 6 as averages over $u$ bins. In Fig. 6, we also include results from the three PAM stations. In total, Fig. 6 represents over 1,200 hourly eddy-correlation measurements of $z_T$ over snow-covered sea ice.

Although the bin-averaged values in Fig. 6 are still somewhat scattered, they are generally within an order of magnitude of what we would predict $z_T$ to be based on (5) and (8). An order of magnitude uncertainty leads to an uncertainty in $C_{HN10}$ of 15–20%. Figure 6, in combination with other recent tests of Andreas’s (1987) $z_T$ model, (8) (e.g., Andreas 2002; Denby and Snellen 2002; Andreas et al. 2003), confirms that the model is a useful tool over surfaces of ice and snow.

4.3 $z_T$ in Summer

We present Fig. 7 as a counterpoint to Fig. 5. This shows the $z_T/z_0$ ratio measured on our main tower during the summer, 15 May–14 September 1998. In contrast to Fig. 6, most of the points in Fig. 7 are above the model prediction. This result implies that the sensible heat fluxes that we measured during the summer were larger in magnitude than what we predict with (1b), (2b), and (8).

We can think of two simple explanations for this mismatch. Either (1b) and (2b) are accurate but (8) underpredicts $z_T$ in summer. Or (8) is still reliable, but something is wrong with (1b) and (2b) during the summer. Because (8) has proved reliable over ice surfaces that were near 0°C (i.e., Denby and Snellen 2002; Andreas 2002, Fig. 3), we lean toward the second explanation.

During the summer, the bare ice surface was near 0°C because of the melting. Surprisingly, though, the water in the melt ponds and leads was significantly warmer than 0°C. For example, Paulson and Pegau (2001) report surface water temperatures in leads in the vicinity of the SHEBA camp of 2°C. In other words, during much of the summer, the surface over which we were sampling
Figure 7. As in Fig. 5, but the zT/z0 ratios here were measured during the SHEBA summer, 15 May–14 September 1998.

did not have a uniform temperature. Consequently, (1b) is not strictly accurate.

We can, however, salvage elements of our bulk flux algorithm using a mosaic technique (e.g., Vihma 1995). We simply estimate the areally averaged heat flux as the area-weighted sum over all surfaces:

\[
H_s = \rho c_p A_i C_{Hi} U_{r,i} (\Theta_{s,i} - \Theta_{r,i}) + \rho c_p A_p C_{Hp} U_{r,p} (\Theta_{s,p} - \Theta_{r,p}) + \rho c_p A_L C_{HrL} U_{r,L} (\Theta_{s,L} - \Theta_{r,L}) .
\]  

(9)

Here, the subscripts i, p, and L refer to conditions over ice, melt ponds, and leads, respectively. Likewise, A_i, A_p, and A_L are the fractional areas of ice, ponds, and leads such that \( A_i + A_p + A_L = 1 \).

We can implement various methods and assumptions for treating (9). But we are not going to discuss these solutions here. Rather, we simply want to explain how (9) is compatible with Fig. 7.

To start, we can assume that the open water areas are small; they therefore do not induce any local circulations. Consequently, \( U_{r,i} \), \( U_{r,p} \), and \( U_{r,L} \) are all approximately the same, as are \( \Theta_{r,i} \), \( \Theta_{r,p} \), and \( \Theta_{r,L} \). Because during the summer everything is near 0°C, the stratification corrections in \( C_{Hi} \), \( C_{Hp} \), and \( C_{HrL} \) are all comparable. Further, if we assume that the roughness lengths \( z_0 \) and \( z_T \) depend only weakly on the surface type, we can reduce (9) to

\[
H_s = \rho c_p C_{Hr} U_r \left[ (A_i \Theta_{s,i} + A_p \Theta_{s,p} + A_L \Theta_{s,L}) - \Theta_r \right] .
\]  

(10)

where \( U_r \) and \( \Theta_r \) are, again, the wind speed and potential temperature at reference height \( r \).

We recognize \( A_i \Theta_{s,i} + A_p \Theta_{s,p} + A_L \Theta_{s,L} \) as the areally averaged surface temperature, \( \Theta_{s,ave} \). Because \( \Theta_{s,p} \) and \( \Theta_{s,L} \) are often higher than \( \Theta_{s,i} \), which is constrained to be 0°C or less, \( \Theta_{s,ave} \) is often higher than \( \Theta_{s,i} \). But we used \( \Theta_{s,i} \) to compute \( z_T \) in Fig. 7. According to (10), this procedure must overestimate \( z_T \) whenever \( \Theta_{s,ave} \) is above \( \Theta_{s,i} \).

In summary, we conclude that Fig. 7 does not mean that (8) is inaccurate during the summer but, instead, that we need to use a mosaic approach to estimate the turbulent heat fluxes when the surface is heterogeneous in temperature. We are still looking for SHEBA data from which we can estimate \( \Theta_{s,ave} \) and, thus, test (10) with our summer data.

5. CONCLUSIONS

The eddy-correlation measurements that we made on our main tower and at three remote Flux-PAM sites during SHEBA provided ample data for developing a preliminary version of the SHEBA bulk flux algorithm. From our yearlong series of drag coefficients, we identified two aerodynamic seasons: winter, when snow was available to blow and drift; and summer, when it was not.

For winter, we parameterize \( z_0 \) with (5), which shows that \( z_0 \) behaves differently in each of three dynamic regimes.

In summer, we recognize that form drag associated with the vertical ice edges surrounding all the open water dominates the air–surface momentum transfer. When we supplement our summer SHEBA data with two other sets that also relate the drag coefficient to ice concentration, we come up with a unified picture of how \( C_{ DN10 } \) depends on ice concentration for the entire range of fractional concentrations, 0 to 1.

For winter, our SHEBA measurements of the roughness length for temperature, \( z_T \), corroborate Andreas’s (1987) model, (8). In summer, the surface was more thermally complex; Andreas’s model tends to underestimate \( z_T \) because we had not accounted for the warm ponds and leads. Our turbulence measurements responded to these, but our calculations of \( z_T \) did not include their influence. We conclude that mosaic calculations will probably be necessary to predict the turbulent heat fluxes correctly in summer.

The values of \( z_0 \) and \( z_T \) that we report have immediate implications for modeling. We find that
$z_0$ and, thus, $C_{Dr}$ are not constant in either winter or summer (see Figs. 1–4). $z_T$ is likewise not constant (see Figs. 5–7). Constancy of the transfer coefficients is a common assumption in large-scale sea ice models (e.g., Holland et al. 1993; Bryan et al. 1996). Figures 5 and 7 also demonstrate that $z_0$ and $z_T$ are not generally equal. From (7), this inequality implies that the transfer coefficients for momentum and sensible heat are generally not equal either. Bryan et al. (1996), for example, assume $z_0 = z_T = 40$ mm. Figures 2 and 6 clearly show that both $z_0$ and $z_T$ over sea ice are one to two orders of magnitude smaller than this value. Finally, comparing Figs. 2 and 6, we see that, generally, $z_0 > z_T$. Equation (7) thus implies that $C_{DN10} > C_{HN10}$. Holland et al. (1993), for example, assume the opposite.

6. ACKNOWLEDGMENTS

The U.S. National Science Foundation supported this work with awards to CRREL, NOAA/ETL, and NPS. The National Science Foundation also supported the work at NCAR. The U.S. Department of the Army provided additional support to Andreas and Jordan through projects at CRREL.

7. REFERENCES


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